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Institute of Polar Studies

Report No. 62

On the Late Precambrian - Early Paleozoic Metavolcanic and Metasedimentary Rocks of the Queen Maud Mountains, Antarctica, and a Comparison With Rocks of Similar Age for Southern Africa

by

Edmund Stump

Institute of Polar Studies
and
Department of Geology and Mineralogy
The Ohio State University

December 1976

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The Ohio State University
Institute of Polar Studies
Columbus, Ohio 43210

ERRATA

Institute of Polar Studies Report Number 62

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Edmund Stump

1. The title should read "On the Late . . . of Similar Age from Southern Africa," not "for Southern Africa."
2. On page 9, the following caption should be inserted:

   Figure 2. Antarctica.
3. On page 11, the following caption should be inserted:

   Figure 3. Queen Maud Mountains, Antarctica.

INSTITUTE OF POLAR STUDIES

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ON THE LATE PRECAMBRIAN - EARLY PALEOZOIC METAVOLCANIC AND METASEDIMENTARY ROCKS OF THE QUEEN MAUD MOUNTAINS, ANTARCTICA, AND A COMPARISON WITH ROCKS OF SIMILAR AGE FOR SOUTHERN AFRICA

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ABSTRACT

The non-intrusive portion of the late Precambrian-early Paleozoic basement complex in the Queen Maud Mountains, Antarctica, consists of graywacke-shale sequences of the Beardmore Group (Goldie, Duncan, and LaGorce Formations), rhyolite porphyries of the Wyatt Formation, and rhyolite porphyries, minor basalts and carbonate, clastic and volcaniclastic metasedimentary rocks of the Taylor and Fairweather Formations.

A model for the evolution of the Ross Orogen between Byrd Glacier and the Pensacola Mountains is presented combining results of this study with published reports from adjacent areas. The deep-sea fan deposits of the Beardmore Group are conformably overlain and intruded by extrusive and hypabyssal portions of the Wyatt Formation. An arc of volcanic islands producing ash-flow tuffs, ash-fall tuffs and lavas, and minor basaltic lavas was active in the early Cambrian. Sediments accumulated in association with the volcanic rocks. All of these rocks were deformed, intruded and metamorphosed to varying degrees during the Ross Orogeny which ended by middle Ordovician.

Rocks of the Damara Orogen of southern Africa were probably linearly continuous with the Ross Orogen prior to the breakup of Gondwanaland. A clastic pulse recognized in late Precambrian sediments in the Cape, the Vanrhynsdorp area and the Nama Basin is interpreted as representing the initiation of the Damara Orogeny in South and South West Africa.

A comparison of the evolution of the Ross and Damara Orogens concludes that the Cape region in South Africa occupied a tectonic regime transitional between the regime of cratonic collision in South West Africa, where orogeny was sustained from the late Precambrian to the middle Orodovician, and the regime of subduction of oceanic crust beneath a continental margin in Antarctica, where two episodes of sedimentation and deformation occurred.

ACKNOWLEDGMENTS

Sincere thanks and appreciation are extended to C.E. Corbató, my adviser, for advice, guidance and criticism, for administrative insight, for field supervision during the 1974-75 season, for an even tone through the ups and the downs, and for a free rein; to D.H. Elliot, for the opportunity to first journey to Antarctica, for continued interest and dialogue on all facets of my research, and for time; to G. Faure, for discussions, for encouragement, and for promotion when it was needed most, at my initiation into the proposal writing game; to K.O. Stanley, for discussions on sedimentary portions of this report; to S.D. Etter and V.P. Wendland, for field assistance in 1970-1971; to P.V. Colbert, for field assistance in 1970-71 and 1974-75; to A.W. Browning, for field assistance in 1974-75, for the quote that starts Chapter 2, and for assorted insights.

To Antarctic Development Squadron Six, U.S. Navy, and in particular to the helicopter pilots and crews at McGregor and Amundsen Glaciers, 1970-71, for excellent field logistics in the Queen Maud Mountains.

In South Africa, to S. Maske, for first awakening an interest in African geology through a course taught during a visiting Professorship at Yale University in 1968-69; to J.F. Enslin and L.E. Kent, for the hospitality of the South African Geological Survey; to H. Jansen, for an excursion through the Waterburg Basin; to K.E.L. Schalk and L.N.J. Engelbrecht, for excursions in central South West Africa; to J.N. Theron, for excursions in the Cape Mountains; to P.G. Sohnge, University of Stellenbosch, for providing maps on the western Cape; to A.E. Schoch, for an excursion to Saldanha Bay; to members of the Precambrian Research Unit, University of Cape Town, including H.J. Blignault, G.J.B. Germs, M.P.A. Jackson, P. Jobert, A. Kroner, V. Vajner, and J.H.W. Ward, for an excursion through their research areas in southern South West Africa and South Africa; to C.J. Hartnady, for an excursion in the Worchester area; to I.C. Rust and H. Geldsetzer, University of Port Elizabeth, for excursions in the Gamtoos area; to the many South African farmers whose lands I crossed, for a collective friendliness unmatched in my travels; to Gerard and Alta Germs and to Hennie Blignault, for hospitality above and beyond the call; to T.W. Gevers, for graciousness, for enthusiasm, and for indefatigable spirit; to Anglo-American Corporation of South Africa, Ltd., for the loan of a field vehicle and a grant of maintenance support from the Group Chairman's Fund, which permitted research on the pre-Cape rocks in Cape Province.

To M.D. Turner, Division of Polar Programs, National Science Foundation, without whose continued support and good faith none of this would have been possible. Research leading to this report has been carried out under National Science Foundation grants GV-26652, GV-33767, and OPP74-04897 to the Institute of Polar Studies and The Ohio State University Research Foundation.

To Peter J. Anderson, for final editing; to Jean Cothran, for typing the production manuscript.

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INTRODUCTION

The principal objectives of this project have been to determine the volcanic and sedimentary environments occurring in the area of the Queen Maud Mountains, Antarctica, during the late Precambrian and early Palaeozoic (Figs. 2 and 3), to gain an understanding of their subsequent tectonism, to relate these environments and events on a regional scale to those determined by other geologists between Byrd Glacier and the Pensacola Mountains (Fig. 9), and finally to compare the late Precambrian-early Paleozoic tectonic development of this portion of Antarctica with that of rocks of similar age in southern Africa, thought to have been laterally continuous prior to the breakup of Gondwanaland.

The initial research was carried out in 1970-71 as part of a helicopter-supported, geological mapping project in the central Transantarctic Mountains. It was my duty to examine as many of the non-crystalline basement exposures as possible in the given time, and localities from Canyon Glacier to the LaGorce Mountains, were visited.

Brief reconnaissance in 1971 had shown that a highly varied stratigraphy and complex structure exist in the Duncan Mountains, and so in 1974-75 a four-man ground party returned to that area to undertake detailed mapping.

The results of these two field seasons and subsequent petrographic studies are presented in the first portion of this report. Appendix A describes each locality examined during the two Antarctic field seasons; descriptions of formations beginning with the stratigraphically oldest with nomenclature following that of previous authors is as follows:

Basement Stratigraphy, Queen Maud Mountains

Canyon-Ramsey Glaciers area	Shackleton Glacier area	Duncan Mountains	Amundsen-Scott Glacier area
		Henson Marble	
	Taylor Fm.	Fairweather Fm.	
	Greenlee Fm.		
			Wyatt Fm.
Goldie Fm.		Duncan Fm.	LaGorce Fm.

In 1972-73 I visited southern Africa in order to examine rocks with ages similar to the Antarctic material to see what comparisons might be made. This was justified by the fact that major orogenic belts of the late Precambrian-early Paleozoic are found both in the Transantarctic Mountains and along the west coast of Africa from the Cape of Good Hope

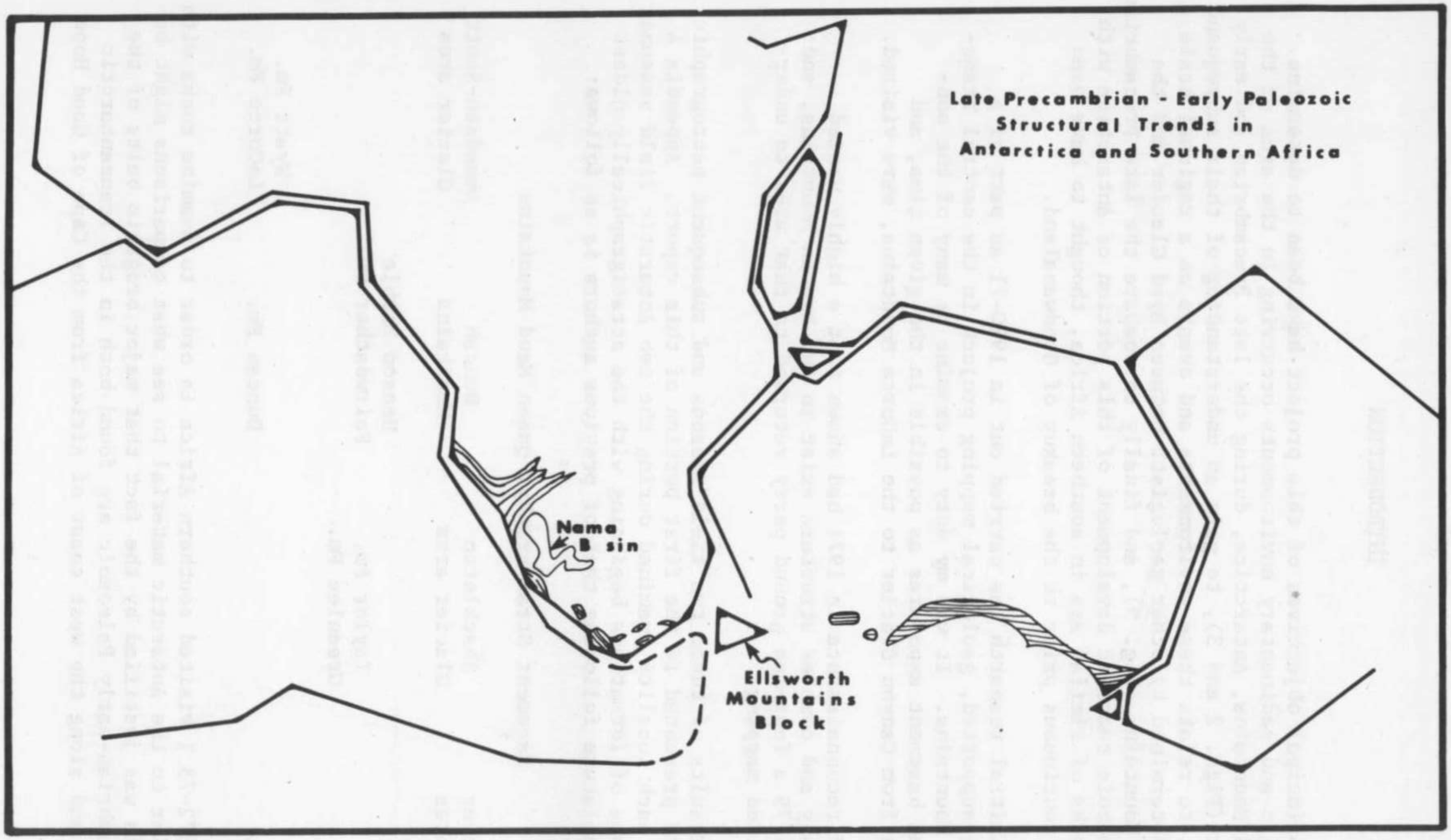


Figure 1. Schematic reconstruction of Gondwanaland, Post-early Paleozoic.

to Gabon, and that on almost all reconstructions of Gondwanaland, Africa and Antarctica are contiguously placed.

Many reconstructions are such that an extension of the late Precambrian-early Paleozoic orogenic trend in Antarctica abuts against much older cratonic rocks of eastern Africa (e.g., Smith and Hallam, 1970), or that the orogenic trend in Africa abuts against older cratonic rocks of East Antarctica (e.g., Tarling, 1972). Both situations seem geologically untenable to me by analogy with the continuous nature of the major mountain systems of the world, such as Tethys and the Cordillera of the Americas. Therefore, I offer my own reconstruction which emphasizes the lateral continuity of the Antarctic and African belts (Figure 1.).

The relative placement of Africa and Antarctica is similar to that of Elliot (1972) and is perhaps best justified by the resultant alignment of latest Permian or early Triassic folds in the Cape Mountains of South Africa and the Pensacola Mountains of Antarctica. The placement of Malagasy in the vicinity of Somalia, Kenya, and Tangania conforms with paleomagnetic data presented by McElhinny and others (1976). Antarctica is fit against Mozambique with removal of post-basement sequences following arguments of Flores (1970). India is placed according to Sclater and Fisher (1974), and Australia according to Weissel and Hayes (1972). The location of the Ellsworth Mountains block is according to Schopf's (1969) hypothesis and is further justified in the final chapter of this report.

For the first two months in Africa I accompanied geologists on excursions in their field areas from central South West Africa to the Cape. The second two months were spent examining the four areas of pre-Cape System rocks east of the major Malmesbury outcrops of the southwestern Cape Province, South Africa (Fig. 14). These areas were chosen for more detailed study because the relationships between them remained poorly understood and because they were closest to Antarctica prior to the breakup of Gondwanaland. Traverses were made across the four windows of pre-Cape rocks following published maps in order to match the nomenclature of previous authors with representative field occurrences, with the expectation that an overview of each area would reveal characteristics common or dissimilar to any or all of them. In addition, samples were collected for petrographic analysis.

The second portion of this report documents these findings and summarizes the geology of South West Africa. Appendix B details the historical development of research on the pre-Cape rocks of the Cape Province, South Africa.

The third and final portion of the report draws conclusions from the tectonic similarities and differences of the late Precambrian-early Paleozoic belts in Antarctica and southern Africa.

Certain geographic names used in the text have been tentatively approved by the Advisory Committee on Antarctic Names of the U.S. Board of Geographic Names. These include:

Mt. Corbató	85°04'S.	165°42'W.
Blackwall Glacier	86°12'S.	159°00'W.
Crack Bluff	86°33'S.	158°38'W.
Wishbone Ridge	84°56'S.	166°56'W.
Mt. Wendland	84°42'S.	175°18'W.
Palid Peak	84°37'S.	178°49'W.
Mt. Ehrenspeck	84°46'S.	175°35'W.

Part I: Antarctica

HISTORICAL BACKGROUND

Early Discoveries

> "Holy shit." This is the kind of sunset you hardly see any more, a 19th-century wilderness sunset, a few of which got set down, approximated, on canvas, landscapes of the American West by artists nobody ever heard of, when the land was still free and the eye innocent, and the presence of the Creator much more direct. Here it thunders now, high and lonely, this anachronism in primal red, in yellow purer than can be found anywhere today, a purity begging to be polluted . . . .
>
> from Gravity's Rainbow
> by Thomas Pynchon

On 28 January 1841 having penetrated to 76^{o} south latitude through pack ice and open water, H.M.S. Erebus and H.M.S. Terror, Capt. James C. Ross, commander, drew near to a twin-peaked volcanic island which ascended 4,000 meters out of the sea and whose higher summit was "emitting flame and smoke in great profusion" (Ross, 1847, p. 216). The island was flanked to the west by a low wall of ice connected to the mainland and to the east-southeast by a vertical ice barrier that rose to 70 meters and extended to the vanishing point.

Here hopes faded of finding an access to the magnetic south pole which already lay to the north behind the high mountain range that for days they had followed to starboard along the east coast of Victoria Land. Ross charted the leading edge of this impassable, floating ice shelf which came to bear his name and then retraced his route to warmer waters, denied the paramount goal of the journey.

After a half century of disinterest the Antarctic again enticed nations to geographical exploration and scientific research. Between 1901 and 1913 three British parties returned with their ships to the smoldering volcano, now called Ross Island, and established bases from which they mounted assaults on the geographic South Pole and undertook scientific investigations on the intervening lands. The basic geological relationships set down by geologists of these parties have served as the foundation for all subsequent research in the Transantarctic Mountains. Ferrar (1907), the geologist on Scott's 1901-04 "Discovery" Expedition, determined that the mountains adjacent to McMurdo Sound are underlain by a metamorphic complex of gneiss, schist and marble, intruded by granite, and overlain by flat-lying arenities which he named the Beacon Sandstone Formation, all of which are intruded by sheets of dolerite.

David and Priestly (1914) of Shackleton's 1907-09 Expedition suggested that the upper Beacon Sandstone belonged to the Gondwana strata occurring throughout the other southern continents, and found that fish plates indicated a Devonian age for beds low in the section. As was customary at the time the gneisses of the basement were assigned to the Archean, but the intrusive granites were thought to be post-Cambrian in age.

This configuration of a truncated basement complex overlain by clastic sedimentary rocks was extended to the Beardmore Glacier area by Shackleton's polar party, who used the glacier as access to the ice plateau, only to be turned back before reaching the pole due to a shortage of rations. On a moraine near the Cloudmaker they collected samples of a limestone breccia that contained Archeocyatha of Cambrian age, and also found in the debris carried by the Beardmore Glacier pieces of greenish-gray slate, supposed to have come from the inaccessible slopes of basement rock bounding the glacier. Mawson (1916) and Skeats (1916) described samples returned by the expedition.

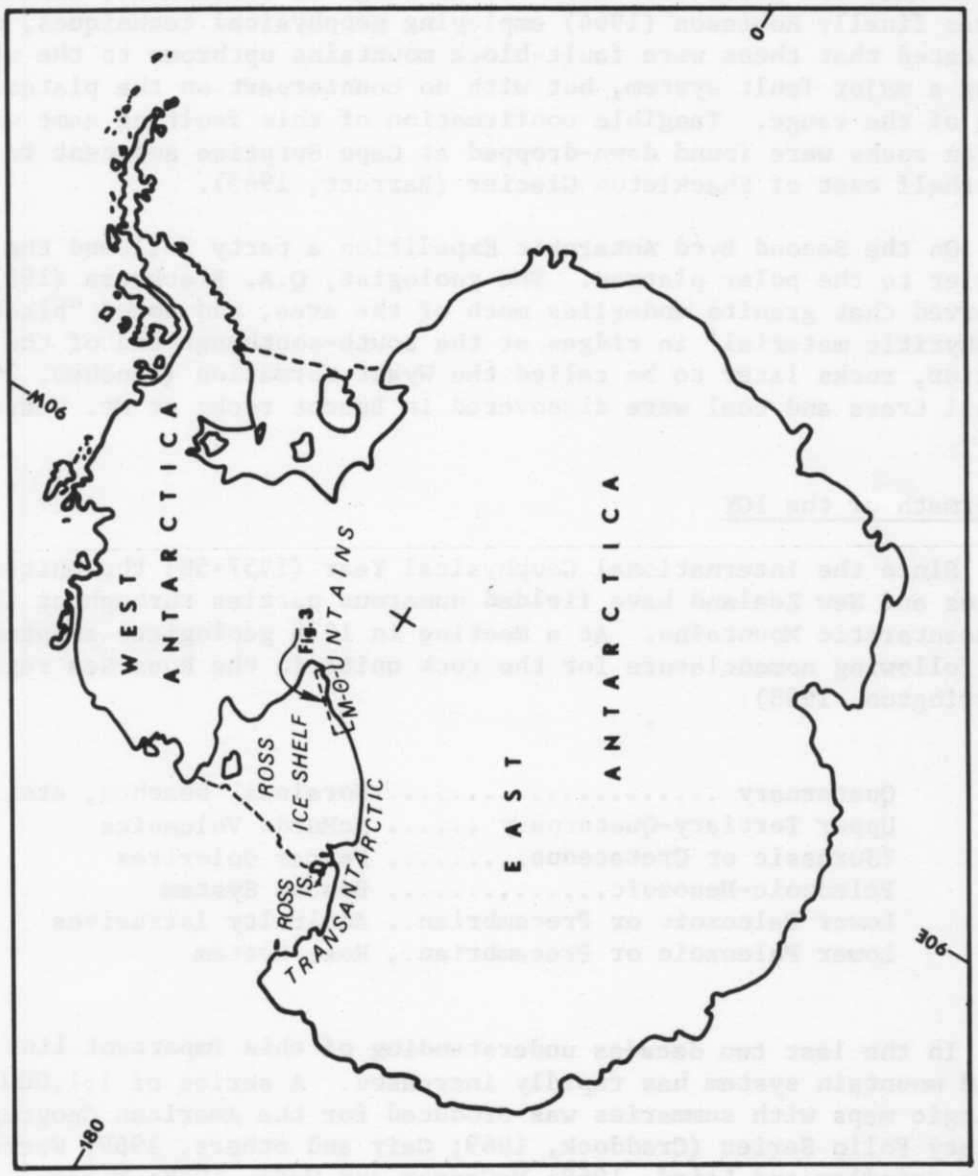
David and Priestly (1914) also set forth the hypothesis that the mountain range was an immense horst, fault-bounded on both sides by downthrown blocks.

Scott's ill-fated "Terra-Nova" Expedition (1910-13) discovered fossilized plant remains of the Permian Glossopteris flora (Seward, 1914) and additional Devonian fish remains (Woodward, 1921). Debenham (1921), the geologist on the expedition, recorded further observations in the McMurdo region, confirming relationships established by his predecessors.

Meanwhile, on their sprint to the pole, the Norwegian party led by Amundsen encountered a continuation of the Transantarctic Mountains at 164° west longitude, traversed them on the Axel Heiberg Glacier and bestowed the name Queen Maud Mountains on that portion of the range from east of the Beardmore Glacier to the eastern horizon. On the return journey about 20 rock samples were collected at Mt. Betty, the only outcrop visited by the party. The samples were described by Schetelig (1915) as orthogeneiss, mica schist, granite and vein quartz. The metamorphic rocks were thought to be Archean or at least younger Precambrian, and the granites, by analogy with the work of David and Priestly (1914), were suggested to be post-Cambrian.

Byrd's Expedition

The next return to this area was by ground parties which travelled between Liv and Leverett Glaciers in support of Byrd's 1928-30 and 1933-35 airborne expeditions. Gould (1931), geologist on the first expedition, recognized that the basement-Beacon division of


WEST ANTARCTICA
EAST ANTARCTICA
ROSS ICE SHELF
ROSS IS.
TRANSANTARCTIC MOUNTAINS
Fig.3
90W
90E
180
0

Victoria Land also applied to these mountains, and collected representative rock samples which were later described by Stewart (1934a; 1934b; 1934c). Gould (1935) also extended David and Priestley's (1914) concept of the great Antarctic horst to the range in this area.

The idea that the Transantarctic Mountains are an immense horst persisted till the 1960's. Hamilton (1960) from work in Taylor Valley, argued that the gross structure of the range is anticlinal; however, it was finally Robinson (1964) employing geophysical techniques, who indicated that these were fault-block mountains upthrown to the west along a major fault system, but with no counterpart on the plateau side of the range. Tangible confirmation of this faulting came when Beacon rocks were found down-dropped at Cape Surprise adjacent to the ice shelf east of Shackleton Glacier (Barrett, 1965).

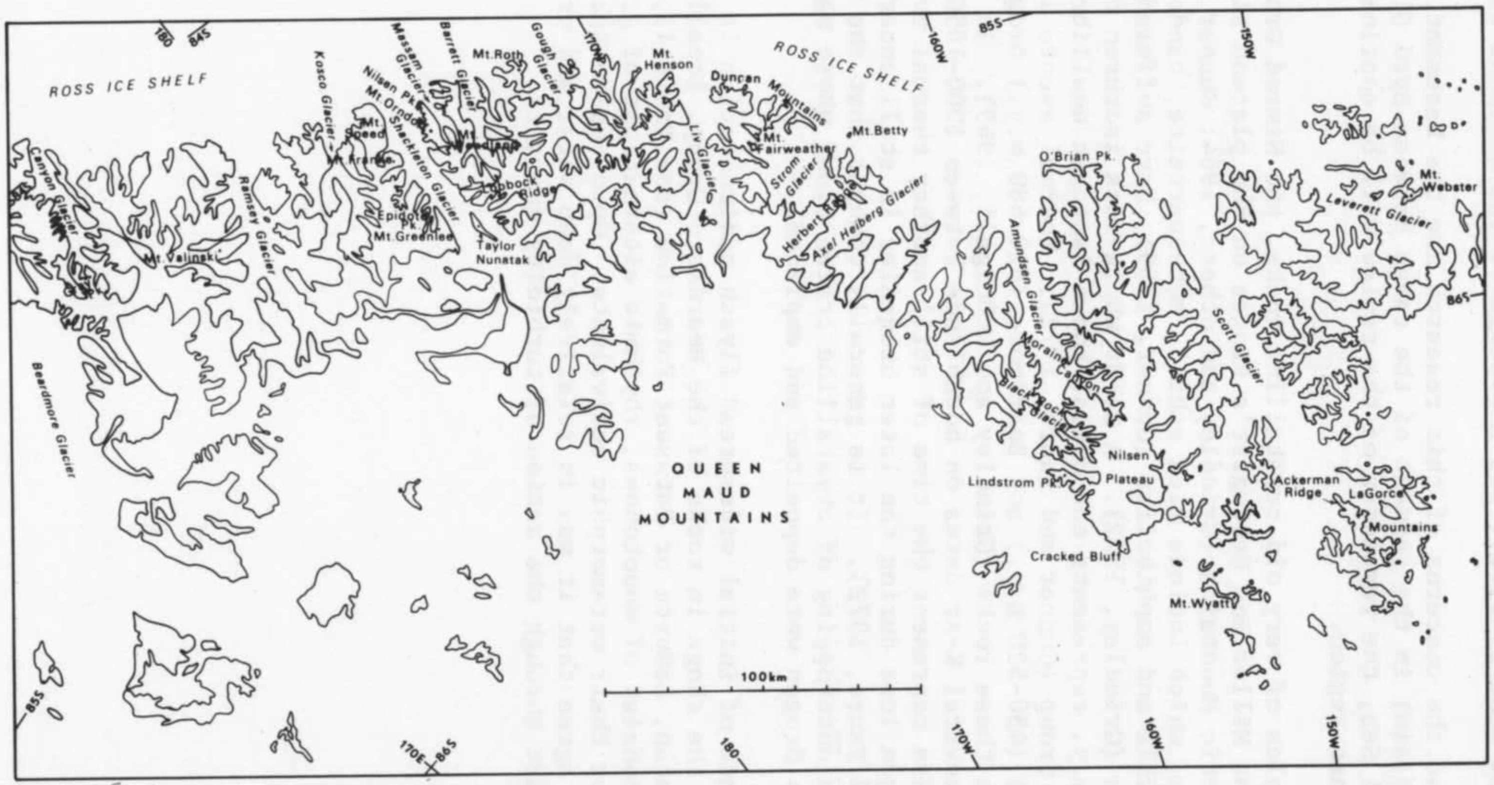
On the Second Byrd Antarctic Expedition a party followed the Scott Glacier to the polar plateau. The geologist, Q.A. Blackburn (1937), observed that granite underlies much of the area, and noted "black, porphyritic material" in ridges at the south-southwest end of the glacier, rocks later to be called the Wyatt Formation (Minshew, 1967). Fossil trees and coal were discovered in Beacon rocks at Mt. Weaver.

Aftermath of the IGY

Since the International Geophysical Year (1957-58) the United States and New Zealand have fielded numerous parties throughout the Transantarctic Mountains. At a meeting in 1958 geologists adapted the following nomenclature for the rock units in the Ross Sea region (Harrington, 1958):

Quaternary	Moraines, beaches, etc.
Upper Tertiary-Quaternary	McMurdo Volcanics
?Jurassic or Cretaceous.........	Ferrar Dolerites
Paleozoic-Mesozoic..............	Beacon System
Lower Paleozoic or Precambrian..	Admiralty Intrusives
Lower Paleozoic or Precambrian..	Ross System

In the last two decades understanding of this important link in the world mountain system has rapidly increased. A series of 1:1,000,000 geologic maps with summaries was produced for the American Geographical Society Folio Series (Craddock, 1969; Gair and others, 1969; Warren, 1969; Grindley and Laird, 1969; McGregor and Wade, 1969; Mirsky, 1969; Schmidt and Ford, 1969). Also, the first four 1:250,000 geological reconnaissance maps of the central Transantarctic Mountains have been produced by the Institute of Polar Studies, The Ohio State University,


ROSS ICE SHELF
ROSS ICE SHELF
QUEEN MAUD MOUNTAINS
Mt. Webster
Leverett Glacier
Scott Glacier
Ackerman Ridge
LaGorce Mountains
Mt. Wyatt
Cracked Bluff
Lindstrom Pk.
Nilsen Plateau
O'Brian Pk.
Amundsen Glacier
Mt. Betty
Duncan Mountains
Mt. Henson
Mt. Fairweather
Strom Glacier
Herbert Range
Axel Heiberg Glacier
Liv Glacier
Gough Glacier
Mt. Roth
Barrett Glacier
Massam Glacier
Nilsen Pk.
Mt. Orndorff
Shackleton Glacier
Mt. Speed
Kosco Glacier
Lubbock Ridge
Epidote Pk.
Mt. Greenlee
Taylor Nunatak
Ramsey Glacier
Mt. Valinski
Canyon Glacier
Beardmore Glacier
100km
150W
160W
170W
180
170E
84S
85S
86S

and the U.S. Geological Survey (Barrett and others, 1970; Lindsey and others, 1973; Barrett and Elliot, 1973; Elliot and others, 1974). Stratigraphic details of the Beacon Supergroup are well documented (Barrett and others, 1972), but knowledge of the basement complex lags.

Because the concerns of this research are the basement rocks (pre-Kukri Peneplain) in the segment of the range between Byrd Glacier and the Weddell Sea, the remainder of the review will be confined to those units in that region.

A complex of very old crystalline rocks, the Nimrod Group, is gound in the Miller and Geologist's Ranges on the plateau side of the Transantarctic Mountains (Grindley and others, 1964; Gunner, 1971a). These rocks, which include mica schist, metaquartzite, banded and augen gneiss, marble and amphibolite (Gunner, 1969), have suffered polyphase deformation (Grindley, 1972). A Rb-Sr whole-rock isochron date of 1984 ± 77 m.y. represents the oldest known isotopic equilibration of the Nimrod Group (Gunner and Faure, 1972). Thermal events imprinted by the Ross (450-520 m.y.) and Beardmore (620-680 m.y.) Orogenies are recorded in these rocks (Grindley and McDougall, 1969). In addition, there are several K-Ar dates on hornblende between 1000-1050 m.y., but whether these represent the time of still another thermal event or partial argon loss during the later orogenies is still uncertain (Gunner and Faure, 1972). It is generally thought that the Nimrod Group is an outcropping of crystalline craton along whose margin rocks of the Ross Orogen were deposited and emplaced.

Evidence of initial widespread flysch sedimentation is found throughout the range in rocks of the Beardmore Group, locally called the Goldie, Duncan, LaGorce or Patuxent Formation (see Table 1). The sequences consist of monotonous, rhythmic alternations of graywacke and shale or their metamorphic equivalents. Opinions on the nature of deposition agree that it was in relatively deep water and probably at least in part through the action of turbidity currents.

Table 1

References to the Beardmore Group

Goldie Formation	LaGorce Formation
Gunn and Walcott, 1962	Doumani and Minshew, 1965
Grindley, 1963	Minshew (1965)
Laird, 1964	Minshew (1966)
Oliver, 1964	Minshew (1967)
Linder and others, 1965	Faure and others, 1968
Wade and others, 1965a	Murtaugh, 1969
Wade and others (1965b)	Katz and Waterhouse, 1970a
LaPrade, 1969	McLelland, unpub.
Gunner, 1971a	
Laird and others, 1971	
Oliver, 1972	
Duncan Formation	**Patuxent Formation**
McGregor, 1965	Schmidt and others, 1964
Linder and others, 1965	Schmidt and others, 1965
	Nelson and others, 1968
	Williams, 1969

Basalt flows, diabase sills, and felsite flows and plugs are interbedded with and intrude the Patuxent Formation in the Neptune Range (Schmidt and others, 1965). A Rb-Sr whole-rock isochron on the felsites indicates emplacement before 1210 ± 76 m.y. (Eastin, 1970), and likewise offers a lower age limit on initiation of sedimentation in that area. Isoclinal folding is characteristic of most areas of exposure. In the Nimrod Glacier area and Pensacola Mountains the folding is truncated by an erosion surface (Laird and others, 1971; Schmidt and others, 1965). But between, in the LaGorce Mountains, the graywacke-shale continues upward uninterrupted into a thick volcanic sequence (Katz and Waterhouse, 1970a). These relationships (unconformities and volcanic extrusion) are cited as evidence of the Beardmore Orogeny (Grindley and McDougall, 1969).

Overlying the unconformity are deposits of limestone which in the Nimrod Glacier area contain Archeocyatha of Early Cambrian age (Laird and Waterhouse, 1962; Hill, 1964; 1965), and in the Patuxent Mountains contain Middle Cambrian trilobites (Palmer and Gatehouse, 1972).

The Shackleton Limestone and associated clastic formations of the Byrd Group occupy an elongate belt from Byrd Glacier to south of Nimrod Glacier (Grindley, 1963; Laird, 1963). The Shackleton Limestone is in part oolitic and contains occasional lenses of conglomerate and sandstone. Locally developed is a 1200 m breccia composed largely of clasts from the underlying Goldie Formation (Laird and others, 1971). Two formations, the Douglas Conglomerate composed of well-rounded pebbles, cobbles and boulders of limestone and quartz, with interspersed sandy and gritty lenses, and the Dick Formation of siltstone with grit lenses, outcrop near Byrd Glacier (Skinner, 1964, 1965). Together their lithologies are similar to the Starshot Formation exposed around Sharshot Glacier, thought to be contemporaneous with the Shackleton Limestone (Laird, 1964). A spilite flow occurring in the Dick Formation and several thin rhyolite flows found in the Starshot Formation indicate minor volcanic activity in the area at this time.

In the Pensacola Mountains, primarily in the Neptune Range, the Nelson Limestone overlies the unconformity with the Patuxent Formation (Schmidt and others, 1964; 1965; Nelson and others, 1968). Basal conglomerates and red-bed clastic rocks, developed locally to a thickness of 20m, are followed by 200-265m of thin-bedded and massive limestone and lesser calcareous shale. These rocks are conformably overlain by the Gambacorta Formation, approximately 100m of rhyolite flows, pyroclastic deposits, volcanic breccia, detrital sandstone, and conglomerate containing mostly volcanic clasts. Intertonguing with and overlying this volcanic unit is the Weins Formation, composed of colorful, thin-bedded shales and sandstones, with several thin, oolitic limestone members (Williams, 1969).

In the intervening region, between the Nimrod Glacier area and the Pensacola Mountains, the transition from the Beardmore to the Byrd Group is at least in part continuous, with the development of voluminous, volcanic sequences. Exposures in the Queen Maud Mountains have remained the poorest known, and are the subject of much of this analysis.

The Thiel Mountains are underlain primarily by massive, porphyritic rhyodacites containing hypersthene and corderite (Ford and Aaron, 1962; Ford, 1964). Indistinct bedding, rare lenticular and eutaxitic structure and profuse broken and embayed, poorly sorted crystals suggest a pyroclastic origin for these rocks (Ford and Sumsion, 1971). Lead-α dates on zircons (600-670 m.y.) indicate a late Precambrian age for the porphyry (Ford and others, 1963; Aaron and Ford, 1964).

Additional massive rhyodacite porphyries crop out in the upper portions of the Reedy, Scott and Amundsen Glaciers, where they are called the Wyatt Formation (Minshew, 1966; 1967). Indirect evidence, including zoned plagioclase and embayed quartz phenocrysts, numerous fractured crystals, abundance of microcrystalline groundmass and lack of flow structures, is again used to suggest that these rocks originated as pyroclastic deposits (Minshew, 1967), although the possibility of a hypabyssal origin is not discounted (Murtaugh, 1969).

A unique occurrence of the Wyatt Formation conformably in contact with the La Gorce Formation has been reported from Ackerman Ridge in the LaGorce Mountains (Katz and Waterhouse, 1970a). There the porphyries are clearly extrusive for they are interbedded with various clastic rocks. However, elsewhere there is no association with sediments. A Rb-Sr whole-rock isochron date of 633 ± 13 m.y. has been reported for the Wyatt Formation from the Wisconsin Range (Faure and others, 1968). Taken together with a Rb-Sr whole rock isochron of 627 ± 22 m.y. on part of the Wisconsin Range Batholith, these dates indicate magma generation and isotopic homogenization in this area during the late Precambrian, activity thought to be related to the Beardmore Orogeny (Grindley and McDougall, 1969).

An isolated section, measured at Mt. Webster north of Leverett Glacier contains several thousand meters of interbedded limestone, sandstone, shale and pyroclastic rhyolite (Minshew, 1967). This is the only exposure which has been visited in a large area of basement rocks between Leverett Glacier and the mouth of Reedy Glacier. Trilobites from a middle portion of the section are Middle(?) Cambrian (Palmer and Gatehouse, 1972) and a Rb-Sr whole-rock isochron on rhyolites higher in the section give a consistent date of 489 ± 30 m.y. (Faure and others, 1968).

The remaining non-intrusive basement outcrops are scattered along the coast of the Ross Ice Shelf from Amundsen to Kosco Glaciers and are particularly well developed on both sides of Shackleton Glacier (Linder and others, 1965; Katz and Waterhouse, 1970b; McGregor, 1965; Wade and others, 1965a, 1965b; LaPrade, 1969). The Fairweather Formation and overlying Henson Marble, occurring in the vicinity of the Duncan Mountains, and the Taylor Formation of the Shackleton Glacier area are known to be composed of felsic volcanic rocks, clastic and calcareous clastic rocks and marbles, but stratigraphic relationships remain obscure (McGregor and Wade, 1969; Wade, 1974).

During the later Cambrian and Ordovician the entire range underwent a severe paroxysm, the Ross Orogeny, which terminated deposition and produced voluminous calc-alkaline magmatism. It appears that all of the rocks of the Ross Supergroup were deformed at this time and that, with the exception of the Pensacola Mountains where folding occurred again during the latest Permian or early Mesozoic (Schmidt and others, 1965), this was the last compressional deformation suffered by the Transantarctic Mountains south and east of the Byrd Glacier.

An extensive batholithic complex with associated migmatites extends almost uninterrupted from approximately Ramsey Glacier to the Horlick Mountains (Burgener, in press; Wade and others, 1965a, 1965b; McGregor, 1965; Murtaugh, 1969; Treves, 1965). To either side, between Byrd and Ramsey Glaciers and in the Thiel and Pensacola Mountains, more isolated plutons cross-cut the Ross Supergroup (Laird, 1964; Gunn and Walcott, 1962; Gunner, 1971a, Ford and Aaron, 1962; Schmidt and others, 1965).

Data from plutons in the Beardmore Glacier area suggest that the magma was generated by partial or nearly complete melting of the Goldie Formation or Nimrod Group into which it is locally intruded (Gunner, 1971a, 1971b, 1974). Contact metamorphism resulting from the plutonism was widespread, but dynamic metamorphic effects were produced only locally (McGregor, 1965). Numerous radiometric dates, derived by K-Ar, Rb-Sr, U-Pb and Pb-α methods, indicate the Ross Orogeny culminated between 450 and 520 m.y. ago (Aaron and Ford, 1964; Craddock and others, 1964; Eastin, 1970; Eastin and Faure, 1972; Faure and others, 1968; Grindley and McDougal, 1969; Gunner, 1971a; Gunner and Faure, 1972; Gunner and Mattinson, 1975; McDougall and Grindley, 1965; Minshew, 1965).

Erosion following this upheaval produced the Kukri Peneplain upon which was deposited the Devonian and younger Beacon Supergroup.

BEARDMORE GROUP

Introduction

The first sediments known to be deposited in the Ross Orogen are the Beardmore Group (Gunn and Walcott, 1962), a suite of graywacke, siltstone and shale, or their metamorphic equivalents, occurring at widely scattered locations throughout the area under consideration. In the western Queen Maud Mountains the Goldie Formation occurs in the central Ramsey Glacier area and around Canyon Glacier. From there it crops out for 300 km in a continuous belt along the ice shelf to north of Nimrod Glacier, the area where it was named the first described (Gunn and Walcott, 1962). At the eastern end of the area the LaGorce Formation crops out in portions of the LaGorce Mountains at the head of Scott Glacier (Minshew, 1967) and a portion of the northwestern flank of Nilsen Plateau. About midway between these occurrences is the Duncan Formation of the Duncan Mountains (McGregor, 1965). Correlation of some or all of these formations has been suggested by various authors (Minshew, 1967; McGregor and Wade, 1969; Wade, 1974), and is supported by work reported here because of field and petrographic similarities.

A late Precambrian age for the Beardmore Group is suggested in the Nimrod Glacier area where the Goldie Formation is unconformably overlain by archeocyathid-bearing, Lower Cambrian Shackleton Limestone (Laird and others, 1971). In addition the LaGorce Formation in the Wisconsin Range is intruded by granitic rocks of the Wisconsin Range Batholith, dated at 627 ± 22 m.y. (Faure and others, 1968).

Goldie Formation

Field Characteristics

The Goldie Formation consists principally of metagraywacke and argillite which exceeds 2500m thickness in the Ramsey Glacier area, provided the section at Mt. Valinski has not been folded. Wade and others (1965a) report a thickness of the Goldie Formation greater than 3300m for the Ramsey Glaicer area. Neither top nor base of the formation is exposed. In outcrop the fresh rock is medium to dark gray, weathering gray to shades of brown. On the western side of Four Ramps the metasedimentary rocks take on a decidedly greenish case and lenses of vein quartz are common.

Bedding is extremely even with bed thickness varying from 0.05-3m but with most beds in the range 0.3-2m. Coarser-grained sediments are found as a rule in the thicker beds. Aside from the beds themselves, lamination is the only common sedimentary structure. However, rare, low-angle cross beds were observed at Mt. Valinski and Gray Peak. Cleavage is commonly developed in the more argillaceous beds and sometimes is difficult to distinguish from the bedding itself.

Plate I - Beardmore Group

Plate I.1 Goldie Formation, Four Ramps. Argillite with biotite grown in conjugate pattern. Crossed nicols, 25x. (ES-107)

Plate I.2 Goldie Formation, Four Ramps. Metagraywacke with clasts of quartz and plagioclase and matrix of quartz, biotite, chlorite and minor calcite. Crossed nicols, 25x. (ES-108)

Plate I.3 Goldie Formation, Mt. Valinski. Calcareous argillite with biotite, quartz and calcite. Crossed nicols, 160x. (ES-142)

Plate I.4 Goldie Formation, Gray Peak. Graywacke with quartz clasts and matrix of biotite with minor calcite and actinolite. Crossed nicols, 25x. (ES-160)

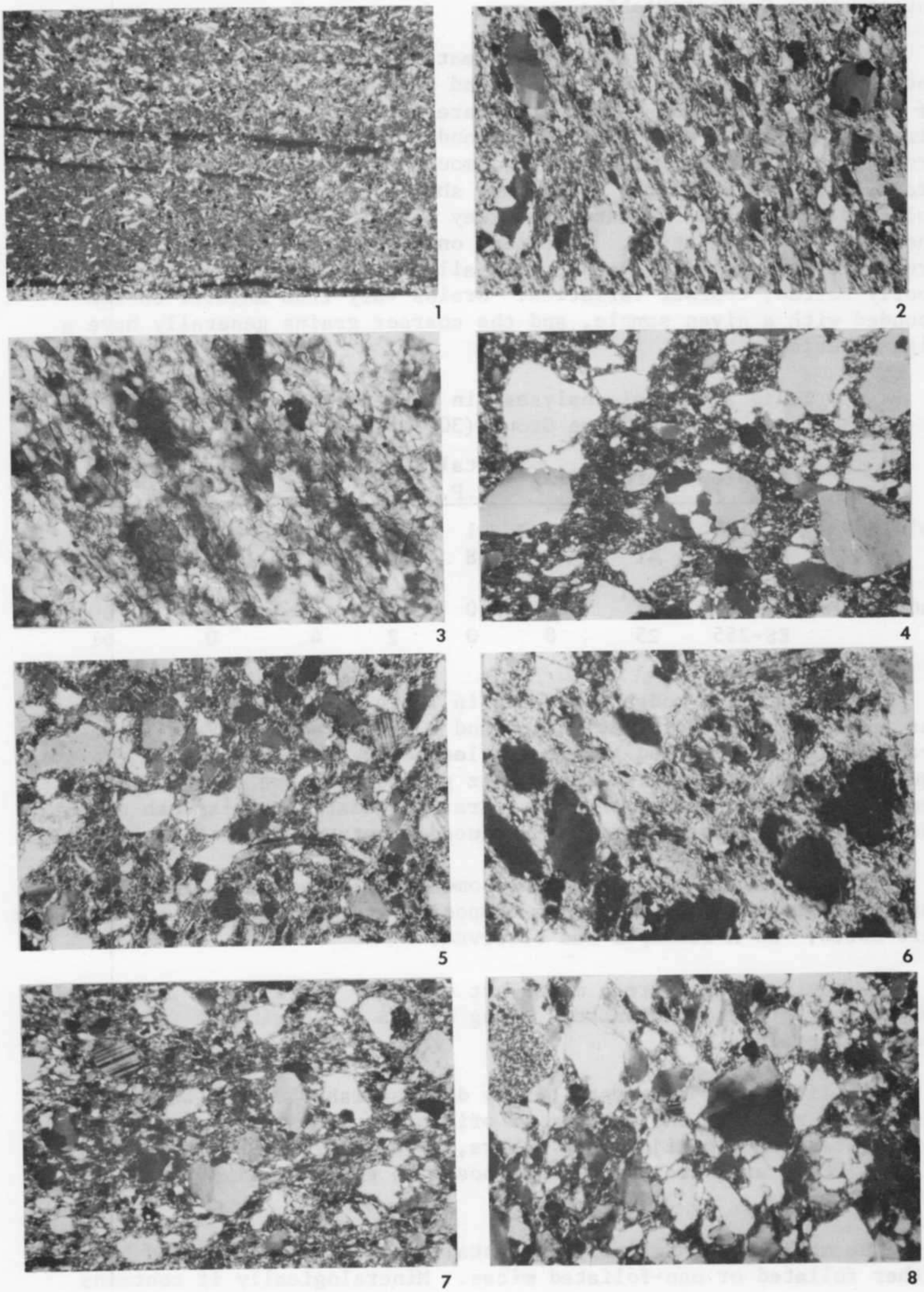
Plate I.5 LaGorce Formation, Black Rock Glacier. Metagraywacke with clasts of quartz, plagioclase and muscovite and matrix of quartz, sericite and chlorite. Crossed nicols, 25x. (ES-230)

Plate I.6 LaGorce Formation, Black Rock Glacier. Close-up of Plate I.5. Crossed nicols, 63x. (ES-230)

Plate I.7 LaGorce Formation, Ackerman Ridge. Metagraywacke with clasts of quartz, plagioclase and muscovite and matrix of quartz and sericite. Crossed nicols, 25x. (ES-293)

Plate I.8 LaGorce Formation, Ackerman Ridge. Metagraywacke with clasts of quartz, plagioclase K-feldspar and rock fragments and matrix of quartz, sericite and chlorite. Crossed nicols, 25x. (ES-299)

PLATE I — BEARDMORE GROUP



Petrographic Characteristics

Petrographically the Goldie Formation is seen to be composed of metamorphosed derivatives of shale and graywacke (see Plate I). Most of the rocks with sand-sized grains are feldspathic graywackes by the classification of Pettijohn, Potter and Siever (1973). Arkosic arenites are represented along with a minor amount of quartz wacke. The sand grains float in a pelitic groundmass showing varying development of foliated or hornfels texture, and they exhibit a crude alignment with the bedding or foliation. Depending on the unit, grain-size may be from silt to very coarse sand, with all sizes being represented in the poorly-sorted, coarser varieties. Grains vary from angular to sub-rounded with a given sample, and the coarser grains generally have a high sphericity.

Table 2. Modal analyses (in percent) of samples from the Beardmore Group (300 counts per sample).

	Sample	Quartz	Plag	Total R.F.	Musc	Opaq	Calcite	Matrix
Goldie Fm.	ES-108	24	6	1	t	t	4	65
	ES-160	41	1	8	t	1	5	44
LaGorce Fm.	ES-230	34	3	0	1	t	0	60
	ES-255	25	8	0	2	4	0	61

Representative modes are shown in Table 2. The quartz usually has a slightly undulose extinction and a few vacuoles. Vaculoes are occasionally aligned and rutile needles occur in a few grains. Equigrannular, polycrystalline quartz grains also are present in small quantities in most samples. Grain boundaries are smooth except where hornfels metamorphism has produced a suturing with the groundmass.

Plagioclase is often twinned, somewhat altered, and finer grained than accompanying quartz. A few composite grains of quartz and feldspar were noted. No K-feldspar was observed.

Muscovite flakes are a minor but visually appreciable fraction of the coarser-grained graywackes, along with a scattering of fine-grained opaque minerals.

Phyllitic rock fragments barely distinguishable from the groundmass occur rarely. The problem of origin of the matrix in graywackes is longstanding (Pettijohn and others, 1973, p. 206), but the rocks of this study area are too metamorphosed to shed any light on the controversy.

The matrix is invariably recrystallized and characterized by either foliated or non-foliated micas. Mineralogically it contains fine-grained quartz and plagioclase with biotite or chlorite or both. Patches of calcite occur in small amounts in most of the coarser-grained samples that were sectioned. Actinolite has been found in one specimen from Gray Peak.

LaGorce Formation

Field Characteristics

The LaGorce Formation was examined in some detail on the northwest flank of Nilsen Plateau, but only briefly in the LaGorce Mountains. Bedding appears to trend approximately parallel to Hansen Spur, the principal occurrence at Nilsen Plateau, and a rough estimate of thickness is that the LaGorce Formation exceeds 2,000m there. Minshew (1967) estimated its thickness in excess of several thousand meters in the LaGorce Mountains, where the section is completely folded (Katz and Waterhouse, 1970a). Neither top nor base is exposed at Nilsen Plateau; however, in the LaGorce Mountains the LaGorce Formation appears to be conformably overlain by the Wyatt Formation (see Appendix A--Ackerman Ridge).

The rocks of the LaGorce Formation are argillites and feldspathic metagraywackes, colored medium to dark gray, some with greenish-gray hues, and weathered to gray or brown.

Petrographic Characteristics

In thin section sand grains can be seen to float in a matrix of recrystallized quartz and sericite or biotite, and in places, associated chlorite (see Plate I). This matrix exceeds 50% of the rock. Clast size varies from silt to coarse sand and is poorly sorted in the coarser-grained units. The angularity varies from sub-rounded coarse grains to angular fine grains. Sphericity is high for the coarse grains. (Modes are given in Table 2).

Quartz extinction is slightly to highly undulose. Vaculoes are usually present in small number and some of the quartz is rutilated. Polygonized quartz grains occur infrequently. Grain boundaries are often raggedly intergrown with the matrix.

Plagioclase is usually twinned and may be slightly altered. Its grain-size is correspondingly smaller than quartz.

Detrital muscovite is particularly conspicuous in samples from Blackwall Glacier, but is also present in rocks from the LaGorce Mountains. Rock fragments were apparent in only a few specimens, although other fragments may be indistinguishable from the matrix.

The matrix material is both foliated and non-foliated. Unlike occurrences in the western portion of the area, no calcite occurs in the matrix of the samples studied from the LaGorce Formation, although Minshew (1967, p. 37) reports secondary veinlets of calcite in rocks from the LaGorce Mountains.

Duncan Formation

Field Characteristics

The Duncan Formation, occurring throughout most of the Duncan Mountains and northeast of Mt. Henson, has undergone metamorphic alteration, to sillimanite grade in places, in which shale and graywacke have been altered to argillite, metagraywacke, schist, and gneiss. Even with the metamorphism bedding characteristics are well preserved in a large portion of the outcrop area.

At its thickest the Duncan Formation exceeds 4600 m with neither top nor base of the formation exposed. Bedding varies from a few cm to 2m and is even where the rocks are unfolded. Lamination is the most common sedimentary structure, occurring in much of the rock.

Dish structures were also observed in beds sporadically distributed throughout the area. They appear in discrete units which are laterally continuous, but end abruptly above and below at bounding beds. The dishes are defined by distinct, dark liminations 0.5-2 mm thick, composed primarily of metamorphic biotite, with spacing generally less than 1 cm. The intervening material is medium- to coarse-grained, polygonized quartz with sutured boundaries, surrounded by a very fine-grained mosaic of quartz, plagioclase, biotite, muscovite and chlorite. The dishes average 10-15 cm in length and are slightly, but distinctly concave upwards. The ends of the dishes terminate either in the intervening quartz-plagioclase material or more commonly against the underside of a higher dish, but nowhere were the laminae seen to cross. Dish structures are thought to be representative of water movement through fluidized sediments (Lowe and LoPiccolo, 1974) and have been reported most often from deep-sea fan deposits.

The single occurrence of a graded-bed, containing a basal conglomerate of sub-angular pelitic clasts was recorded on the west prong of Wishbone Ridge. The conglomerate fragments range from less than 1 cm to 10 cm, with most in the smaller size ranges. Matrix constitutes more than 50% of the rock and separates most of the clasts from each other. The conglomerate is up to 10 cm thick but is discontinuous horizontally. It has a sharp contact with the underlying bed and grades upwards into metagraywacke.

This graded bed is the only one which I observed anywhere in the Queen Maud Mountains; however, this may have been due to alteration of the rocks or to the reconnaissance nature of the study. Graded-bedding is apparently rather common elsewhere in the late Precambrian graywacke-shale deposits of the Transantarctic Mountains for numerous other authors have reported occurrences: Robertson Bay Group: Harrington and others, (1967); Goldie Formation: Laird and others, (1971); Gunner (1971a); Duncan Formation: McGregor (1965); LaGorce Formation: Minshew (1967); Murtaugh (1969); Patuxent Formation: Schmidt and others (1964). Sole markings and flute casts are reported from the LaGorce Formation by Minshew (1967) and Murtaugh (1969) and from the Patuxent Formation by Schmidt and others, (1964).

Petrographic Characteristics

Petrographically the rocks appear to retain little of their original character except for remnants of quartz grains in the coarser metagraywackes. A characteristic brown, green, and white layered rock is scattered throughout the section. Calc-silicate assemblages of tremolite/actinolite-biotite-quartz-K-feldspar-plagioclase-sphene indicate that the original sediments were calcareous in part.

Environment of Deposition

It is concluded that the Goldie, Duncan and LaGorce Formations represent deep-sea fan deposits which accumulated through the action of turbidity flows. Features of these rocks indicative of such an environment include: (1) very regular bedding, with alternating coarse- and fine-grained layers, (2) presence of fine lamination, massive and graded beds and (3) the general immaturity of the sediments (see Walker and Mutti, 1973).

Using Bouma's (1962) notation, combinations of a, b, d, and e units are represented in these formations. Unit c, represented by wavy or convolute laminations or ripples, apparently is seldom developed in these rocks. However, Laird and others (1971) have recorded "ripple-drift bedding" at two localities of the Goldie Formation in the Nimrod Glacier area.

WYATT FORMATION

Introduction

The Wyatt Formation, found adjacent to the polar ice sheet from Nilsen Plateau to Reedy Glacier, is a massive, silicic, porphyritic felsite. It appears to conformably overlie the LaGorce Formation on Ackerman Ridge in the LaGorce Mountains, where it occurs as extrusive volcanic units interbedded with clastic metasediments (see Appendix A, Ackerman Ridge). Elsewhere the rock is devoid of primary, internal structure, although a foliation is developed in many areas due to post-emplacement deformation.

Minshew (1967) named the formation for outcrops at the head of Scott Glacier, including Mt. Wyatt, the informal type locality, and Murtaugh (1969) later studied other occurrences at the head of Reedy Glacier. The conformable contact on Ackerman Ridge was discovered by Katz and Waterhouse (1970a). McLelland (unpub.) examined basement exposures on the southwestern flank of Nilsen Plateau in 1963-64, but his results have remained unpublished. In 1971 I worked on much of the outcrop area of the Wyatt Formation on the western and southwestern flank of Nilsen Plateau, and spent one day on Ackerman Ridge, where cross-beds in a clastic portion of the Wyatt Formation indicate that it lies stratigraphically above the LaGorce Formation.

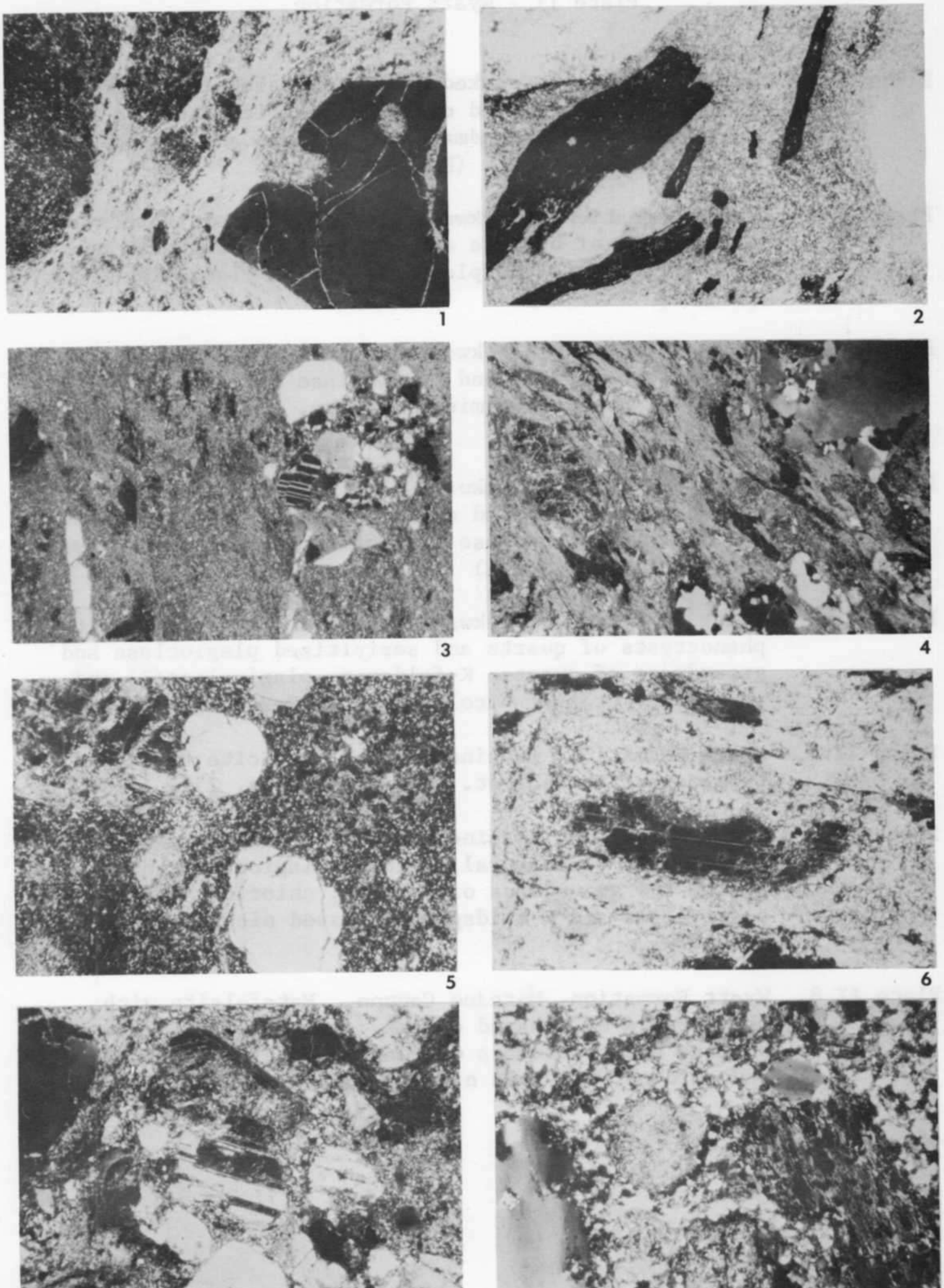
Petrographic Characteristics

Where examined, the Wyatt Formation has a black, dark gray or orange groundmass, sprinkled throughout with quartz, plagioclase, biotite and sometimes orthoclase phenocrysts, as well as basic rock fragments. Petrographically the quartz is seen to be either rounded or subhedral, and often embayed (see Plate II). Broken fragments also occur in most samples. The plagioclase is subhedral to euhedral and is twinned. It is always to some degree altered to sericite and in many specimens completely replaced by it. In several samples from northwest of Lindstrøm Pk. the plagioclase has been saussuritized. In most samples the percentage of plagioclase exceeds that of quartz (Table 3). Subhedral to euhedral orthoclase occurs in samples from west of Mt. Sundbeck and northwest of Lindstrøm Pk. but is absent in the great bulk of the Wyatt Formation underlying the Moraine Canyon-Blackwall Glacier area.

Table 3. Modal analyses (in percent) of samples from the Wyatt Formation (300 counts per sample).

Sample	Quartz	Plag	K-spar	Biot	V.R.F.	Opaq	Calcite	G-mass
ES-208	11	14	0	0	1	0	0	74
ES-217	13	22	4	0	0	0	0	61
ES-227	8	17	t	3	2	1	0	69
ES-228	6	14	0	2	0	2	1	75
ES-240	11	27	0	0	2	2	1	57
ES-247	16	9	0	1	3	1	0	70
ES-249	5	30	0	0	0	1	0	64
ES-251	11	36	t	0	0	1	0	52
ES-274	7	2	0	0	0	0	0	91
ES-290	10	15	0	0	2	t	2	71
ES-301	13	21	0	0	3	2	0	61

PLATE II — WYATT FORMATION



1 2 3 4 5 6 7 8

Plate II - Wyatt Formation

Plate II.1 Wyatt Formation, Cracked Bluff. Metafelsite with phenocrysts of embayed quartz and sericitized plagioclase and groundmass of sericite and quartz. Crossed nicols, 25x. (ES-208)

Plate II.2 Wyatt Formation, Blackwall Glacier. Metafelsite with phenocrysts of biotite and quartz and groundmass of sericite, quartz and plagioclase(?). Plane light, 25x. (ES-227)

Plate II.3 Wyatt Formation, Blackwall Glacier. Metafelsite with fragments of quartz and plagioclase and groundmass of sericite, quartz and minor biotite. Crossed nicols, 25x. (ES-232)

Plate II.4 Wyatt Formation, Blackwall Glacier. Metafelsite with phenocrysts of embayed quartz and altered plagioclase and foliated groundmass of biotite and quartz. Crossed nicols, 25x. (ES-234)

Plate II.5 Wyatt Formation, Blackwall Glacier. Metafelsite with phenocrysts of quartz and sericitized plagioclase and groundmass of quartz, K-feldspar, plagioclase(?) and sericite. Crossed nicols, 25x. (ES-243)

Plate II.6 Wyatt Formation, Moraine Canyon. Sericite replacing plagioclase phenocryst. Crossed nicols, 25x. (ES-307)

Plate II.7 Wyatt Formation, Moraine Canyon. Metafelsite with phenocrysts of euhedral, twinned plagioclase and rounded quartz and groundmass of biotite, chlorite, quartz, plagioclase and K-feldspar. Crossed nicols, 25x. (ES-307)

Plate II.8 Wyatt Formation, Moraine Canyon. Metafelsite with phenocrysts of embayed quartz, altered plagioclase and biotite and groundmass of quartz, K-feldspar, chlorite and epidote. Crossed nicols, 25x. (ES-308)

Phenocrysts of altered, platy biotite (up to 3 mm in size) occur in all samples examined except one from northwest of Lindstrøm Pk. and two from Ackerman Ridge. By contrast, no phenocrysts of mica were found in any of the volcanic rocks from the Taylor or Fairweather Formations to the northwest. The presence of biotite appears to be a useful criterion for distinguishing the Wyatt Formation from the other porphyritic felsites in the region. Altered biotite always occurs to some degree as opaque minerals ghosting the original platy form of the phenocrysts.

Patches of finely crystallized chlorite or biotite, sometimes associated with opaque grains, are found in most of the samples. The discrete relationship of these bodies to the mesostasis suggests that they are altered basic rock fragments rather than being derived by some process of alteration of the groundmass.

The appearance of the groundmass of the Wyatt Formation shows considerable variation areally. On the spurs at the head of Blackwall Glacier where the LaGorce and Wyatt Formation are in contact, it is a massive intergrowth of extremely fine-grained (< 0.03 mm) quartz, feldspar(?), and sericite, biotite and chlorite. Elsewhere the mineralogy is similar although amounts of specific minerals change. K-feldspar may or may not be present and usually only one mica occurs at a time. Hematite is conspicuous in one sample from Moraine Canyon.

The texture of the groundmass shows considerable variation. Grain size ranges from extreme fineness at the head of Blackwall Glacier to a maximum (0.125 mm, average) in portions of Moraine Canyon. There appears to be greater alteration of feldspar in the coarser-grained specimens. Fine-grained micas are dispersed between and sometimes within the feldspar grains, and may be segregated into mottled, filamentous, or strongly foliated patterns. The assemblage biotite-tremolite-epidote represents the highest metamorphic grade attained in these rocks.

The groundmass of one specimen differs from the rest, and is thought to represent a primary igneous texture produced by a cooling rate slower than that of the surrounding rock. The texture is hypidiomorphic-granular with slightly altered, subhedral plagioclase intergrown with granular quartz. Chlorite and opaque minerals occur in coarse-grained segregated patches, but show little dispersion throughout the quartz and feldspar.

Nature of Emplacement

Minshew (1967) stated that the Wyatt Formation is volcanic rock. Murtaugh (1969) suggested either extrusive or hypabyssal emplacement. Both authors pointed to the fragmental and embayed nature of the phenocrysts to suggest that these rocks were of pyroclastic origin, but no positive textural evidence was presented. Although rocks of the Taylor Formation display.welded shard structure, spherulites and other characteristics of volcanic and pyroclastic rocks, no similar features have been found in any of the samples from the Wyatt Formation. These structures, if originally present, could have been totally obliterated during the Ross Orogeny, as is thought to have happened to the Fairweather Formation.

The extrusive origin of the Wyatt Formation is confirmed on Ackerman Ridge where it is interbedded with clastic sediments, and northwest of Lindstrøm Peak where bedding is indicated by concentrations of phenocrysts in faint layers and by thin (1 cm) lenses of "chert" at one location. However, the relationships around Blackwall Glacier show that the Wyatt Formation is at least in part intrusive there. The LaGorce Formation exposed on the western and southern walls of the canyon appears to be folded into a large N-S trending syncline whose eastern limb is truncated at a high angle on three spurs by the Wyatt Formation. Fragments of the LaGorce graywacke are contained in the Wyatt porphyry. No baked zone is present in the LaGorce Formation there, nor is a chilled zone developed in the Wyatt, which exhibits a featureless micro-crystalline groundmass throughout its exposure, except for the development of foliation in several wide shear zones. For this contact to be an unconformity, with extruded Wyatt Formation overlying eroded LaGorce Formation requires a two-phase folding history, a situation not indicated by field evidence.

Felsites on a ridge crest west of Mt. Sundbeck clearly intrude graywackes there. This occurrence of the Wyatt Formation differs from others in the Moraine Canyon-Blackwall Glacier area in that it contains clear, rose-pink quartz phenocrysts, in general lacks biotite phenocrysts, and has composites of quartz grains to 15 cm in diameter. At another location near the head of Blackwall Glacier a nearly horizontal section of cross-bedded quartzite with a 1m thick basal bed of highly sheared white and green marble overlies the Wyatt Formation. The nature of the contact, whether sedimentary, structural or intrusive, is again equivocal.

To recapitulate, there is no evidence at hand in the Moraine Canyon-Blackwall Glacier area to confirm extrusive relationships and ample evidence rather to suggest a hypabyssal origin for the Wyatt Formation there.

TAYLOR FORMATION

Introduction

The Taylor Formation crops out on both sides of the northern portion of Shackleton Glacier. Taylor Nunatak was deisgnated the informal type locality by Wade and others (1965, 1965b), who named and first studied the formation, reporting "a thick unit comprised of quartzites, conglomerates, calcareous quartzites, and marbles." Later felsic volcanic rocks were also recognized (McGregor and Wade, 1969; Wade, 1974; Stump, 1974). The fine-grained metasedimentary rocks underlying the lowermost metavolcanic rocks at Mt. Greenlee and Epidote Peak were also later designated the Greenlee Formation (Wade, 1974).

A tract of gneiss and granite occurs in the northwest portion of the area. McGregor and Wade's (1969) map pictures the gneisses as Taylor Formation and as granite. Mapping by our 1970-71 field party found it useful to dintinguish the Franke Migmatites from the Taylor Formation though it is likely that at least some of the migmatites were derived from the less deformed and metamorphosed Taylor Formation. The reader may refer to Burgener (1975) for descriptions of the gneisses, which are outside the scope of this report.

Field Occurrences

Three distinct lithologic associations are found in specific portions of the lower Shackleton Glacier area. Outcrops on the west side of Shackleton Glacier include basic to silicic volcanic rocks and fine-grained, micaceous quartzites. The quartzites, which occur on the northern end of Mt. Greenlee and the lower slopes of Epidote Peak are dark gray to brown and evenly bedded at 5 cm to 1 m intervals. These pass upward into a series of basaltic flows, some of which are amygdaloidal and some of which are brecciated. The section continues on the ridge crest west of Epidote Peak where highly sheared silicic volcanic rocks are found. These massive rocks are shades of brown and gray and may be porphyritic or non-porphyritic. At the west end of the ridge several units of white, coarsely crystalline marble occur.

At Taylor Nunatak a massive body of gray to brown porphyritic felsite is overlain by a section of limestones and volcaniclastic rocks interbedded with lavas and other acidic volcanic rocks (see Appendix A--Taylor Nunatak for measured section). The limestones are white or light gray and may contain thin rhythmic intercalations of "chert," which may be continuous or discontinuous. These carbonate units are interbedded with coarse-grained volcaniclastic rocks in which bedding may be either distinct or indistinct.

In the remainder of the area on the east side of Shackleton Glacier the rocks are rhyolites interbedded with medium- to coarse-grained, cross-bedded quartzites. The rhyolites occur in individual units ranging between 2 m and 100 m and appear not to have internal subdivisions. The rocks are massive, gray or tan, and invariably contain phenocrysts of quartz and feldspar. The quartzites occurring between the volcanic units are light gray to white and contain ubiquitous festoon cross-beds.

Field relations suggest that the oldest rocks occur on the west side of Shackleton Glacier, the youngest on the east side, and an intermediate section at Taylor Nunatak (arguments for these relationships are found in Appendix A--Shackleton Glacier area).

The volcanic rocks of the Taylor Formation will be described first, beginning with units lower in the section, and descriptions of the sedimentary rocks will follow, again arranged with the lower units first.

Volcanic Rocks

Basaltic Rocks

The oldest volcanic deposits in the Taylor Formation, occurring immediately above the Greenlee Formation at Epidote Peak and Mt. Greenlee, are altered basaltic lavas, some of which are amygdaloidal (see page 57). These are interbedded with metagraywacks and phyllite similar to the sedimentary rocks of the Greenlee Formation, and give way higher in the section on the ridge west of Epidote Peak to more silicic felsites which are highly deformed.

A typical sample from Mt. Greenlee contains altered, bladed plagioclase laths separated by secondary actinolite, sphene, a little quartz, and patches of chlorite. The texture is intergranular with development of secondary ferromagnesian minerals. Also on Mt. Greenlee can be found an amygdaloidal lava with plagioclase phenocrysts to 3 mm and aligned, sparsely developed microlites (Plate III.1). The groundmass is a dark mesostasis of finely intergrown actinolite needles, plagioclase and granular epidote. Again chlorite aggregates occur infrequently.

A basaltic breccia crops out at the upper northern portion of Mt. Greenlee (Plate III.2). The fragments are ragged and aligned with sizes ranging from 0.1-2.5 cm. They contain sparse, aligned plagioclase microlites in a groundmass of actinolite, biotite, plagioclase, epidote and sphene, and all fragments within the breccia appear to have been derived from the same source. The basalt chips are surrounded by sparry calcite or sparry calcite, epidote, zoisite and actinolite. Close packing and alignment of the fragments indicate probable subaqueous deposition, in contrast to the hcaotic breccias usually formed by terrestrial processes.

PLATE III — TAYLOR FORMATION, VOLCANIC ROCKS

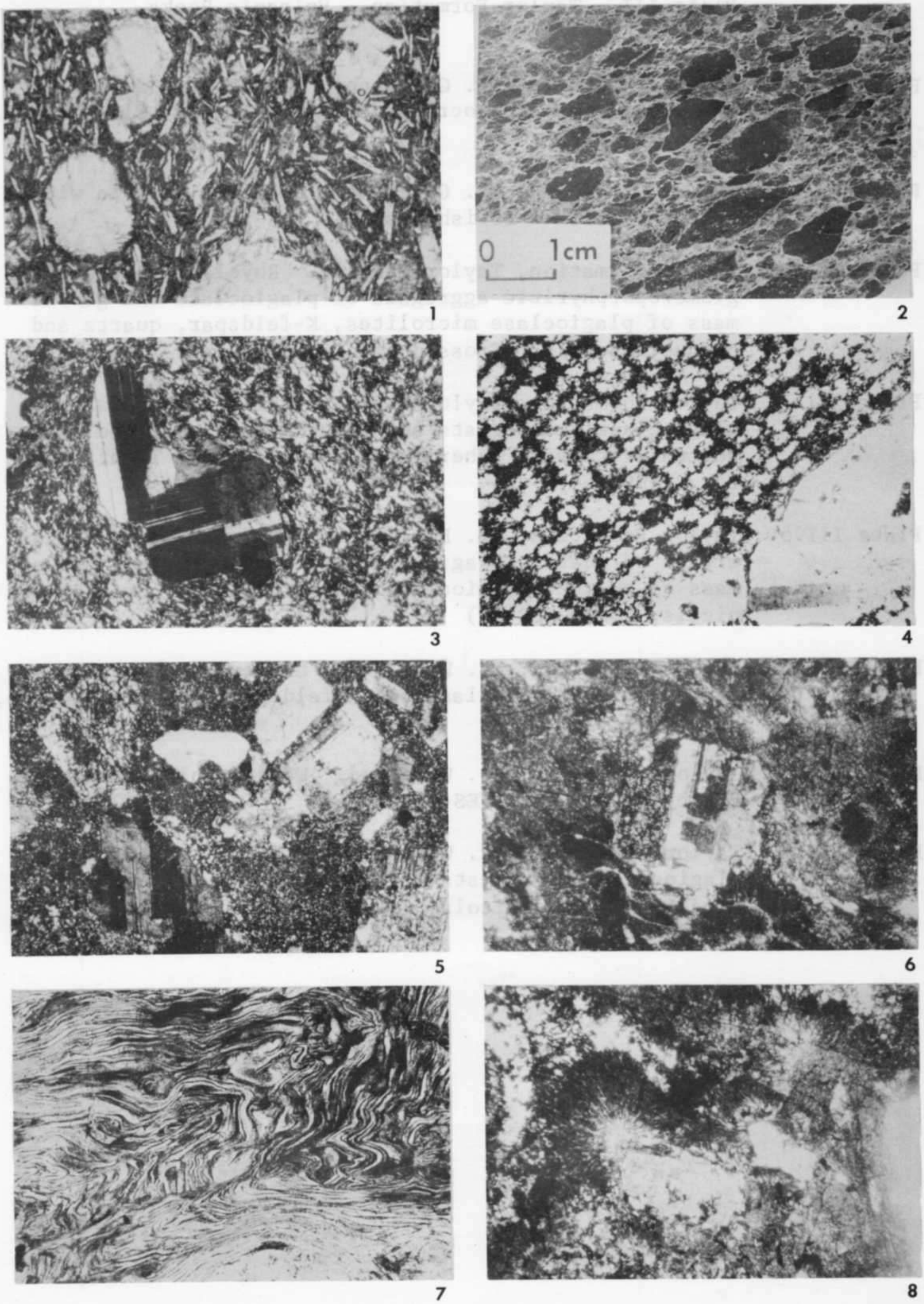


Plate III - Taylor Formation - Volcanic Rocks

Plate III.1 Taylor Formation, Mt. Greenlee. Basalt with amygdule and plagioclase phenocrysts and microlites. Plane light, 25x. (ES-96)

Plate III.2 Taylor Formation, Mt. Greenlee. Basaltic breccia with calcite matrix. Polished slab. (ES-95)

Plate III.3 Taylor Formation, Taylor Nunatak. Rhyolite lava with glomeroporphyritic aggregate of plagioclase and groundmass of plagioclase microlites, K-feldspar, quartz and opaque minerals. Crossed nicols, 25x. (ES-193)

Plate III.4 Taylor Formation, Taylor Nunatak. Silicic lava with plagioclase phenocrysts and groundmass of plagioclase and chlorite with spherical texture. Plane light, 25x. (ES-184)

Plate III.5 Taylor Formation, Mt. Ehrenspeck. Rhyolite with phenocrysts of quartz, plagioclase and K-feldspar and groundmass of quartz, plagioclase and K-feldspar. Crossed nicols, 25x. (ES-40)

Plate III.6 Taylor Formation, Mt. Wendland. Glomeroporphyritic aggregate of plagioclase and K-feldspar. Crossed nicols, 25x. (ES-61)

Plate III.7 Taylor Formation, Mt. Wendland. Welded shard structure. Plane light, 63x. (ES-61)

Plate III.8 Taylor Formation, Mt. Wendland. Felsite containing plagioclase phenocrysts with spherulites and perlitic cracks. Crossed nicols, 25x. (Es-61)

Silicic Lavas

Silicic lavas are found at Epidote Peak, north of Mt. Orndorf, and in several units at Taylor Nunatak-North. Except for the occurrence at Mt. Orndorf all examples contain euhedral plagioclase phenocrysts, sometimes a glomeroporphyritic aggregates (Plate III.3). The groundmass varies, but is most commonly a pilotaxitic or granular intergrowth of plagioclase plus ferromagnesian and other minerals, including biotite, actinolite, epidote, magnetite and quartz. Microlites are well developed in samples from Epidote Peak. Samples from Taylor Nunatrak-North show microlites or plagioclase grains in randomly oriented anhedral or radiating forms. An unusual groundmass texture occurs in one sample from Taylor Nunatak-North where small (0.5 mm) spheres of microcrystalline plagioclase are separated by a network of very fine chlorite (Plate III.4).

A sample from north of Mt. Orndorf contains a granular intergrowth of plagioclase microlites, quartz, chlorite, epidote and sphene. Chloritic patches representing altered basic rock fragments are common and amygdales also occur.

Silicic Felsites

The most widespread volcanic rocks of the Taylor Formation are prophyritic felsites of rhyolitic composition, which occur at numerous outcrops, particularly on the east side of Shackleton Glacier (see page 57). They are characteristically dark gray to black, less often light gray and tan, with numerous phenocrysts of white feldspar and clear quartz. Where interbedded with sediments, these felsites range from 2 m up to 100 m in thickness with no discernable beds within a given unit. At some localities adjacent to the glacier, massive bodies of felsite comprise still wider areas of outcrop.

Petrographically these rocks are seen to be composed of phenocrysts of quartz, plagioclase and in some cases orthoclase set in a microcrystalline groundmass (Plate III.5). Plagioclase predominates over quartz in more than three-fourths of the point-counted samples and in only two samples does orthoclase exceed ten percent of the total phenocrysts (see Table 4).

Accompanied by small, angular chips of quartz and feldspar, the larger phenocrysts and rock fragments are usually unsorted and randomly distributed. The quartz is commonly embayed and fractured and may be euhedral, subhedral, or anhedral. Plagioclase often has albite or Carlsbad twins, and is sometimes zoned. It is usually euhedral, occasionally embayed, and may be highly altered to sericite or epidote. In a few specimens the phenocrysts are glomeroporphyritic aggregates of several crystals. Compositions of plagioclase

Table 4. Modal analyses (in percent) of volcanic samples from the Taylor Formation (300 point counts per sample).

Sample	Quartz	Plag	K-spar	V.R.F.	Opaq	Calcite	G-mass	An%
ES-13	11	15	0	0	0	0	74	4
ES-19	10	12	0	2	t	t	76	
ES-21	17	17	1	t	1	0	64	
ES-38	15	11	15	0	1	0	59	
ES-40	4	28	2	6	1	0	59	
ES-42	14	10	18	1	0	0	57	
ES-45	4	15	1	0	1	0	79	
ES-52	8	6	0	0	0	0	86	35
ES-53	4	6	0	0	1	0	89	
ES-61	11	6	2	4	t	0	77	31
ES-130	15	8	0	7	t	0	70	
ES-136	17	8	0	1	0	0	74	34
ES-186	t	4	0	0	1	0	95	34
ES-190	0	7	0	0	0	0	93	32
ES-193	2	5	0	2	0	0	91	35
ES-201	4	5	0	0	t	3	88	26

fall in the oligoclase to andesine range (An_{26-35}) (Michel-Levy method). One sample contains numerous glomeroporphyritic aggregates of plagioclase and orthoclase (Plate III.6). Orthoclase phenocrysts are subhedral to euhedral and show little sign of alteration. Volcanic rock fragments, composing a small portion of many samples, may be silicic in composition, although the majority are more basic types.

The groundmass of most of the felsites examined petrographically is a featureless, partially recrystalized mesostasis. As a rule crystals in the matrix show no preferred orientation; however zones of foliation are developed along the northeastern margin of the outcrop area, toward the belt of mylonite at Nilsen Peak. Discernible minerals in the groundmass include quartz, K-feldspar and plagioclase, as well as sericite and fine opaque minerals. Isotropic portions of the groundmass are not uncommon in samples where recrystallization is slight. Upon staining these areas generally show the presence of potassium. It is likely that primary textural features have been obliterated by recrystallization in many of these rocks. A few samples do, however, preserve original textures, which are most useful for interpretation of the eruptive mode of these porphyries.

One sample from near Mt. Wendland preserves an excellent example of welded shard structure (Plate III.7). Individual shards are compressed and in parts are highly contorted. Spherulites of quartz and feldspar radiate through much of the groundmass and axiolitic structure is preserved in some of the devitrified shards.

Another sample from the same area is similar in appearance, the groundmass exhibiting fine, subparallel, discontinuous lines, again interpreted as highly compressed welded shard structure (Plate IV.3). Numerous spherulites have grown across the layering and in a few cases displaced it slightly, indicating that devitrification began before the rock was completely solidified.

A sample from Lubbock Ridge contains numerous bubblewall shards, many displaying a tricuspate form (Plate IV.4). The shards are undistroted and are encased in a matrix clouded with finely disseminated opaque grains. One noticeable plagioclase phenocryst is still surrounded by a cuspate border of bubble walls (Plate IV.5). Pumice fragments (to 2 mm) are common throughout and volcanic rock fragments also occur. Recrystallization of the shards, pumice and groundmass is slight, although axiolites are developed in some of the shards. Sodium-cobaltinitrate preferentially stains the shards and pumice making them stand out but obscuring their internal structures. This sample was collected from the uppermost portion of its volcanic unit, probably accounting for the unwelded nature of the shards.

PLATE IV — TAYLOR FORMATION, VOLCANIC ROCKS

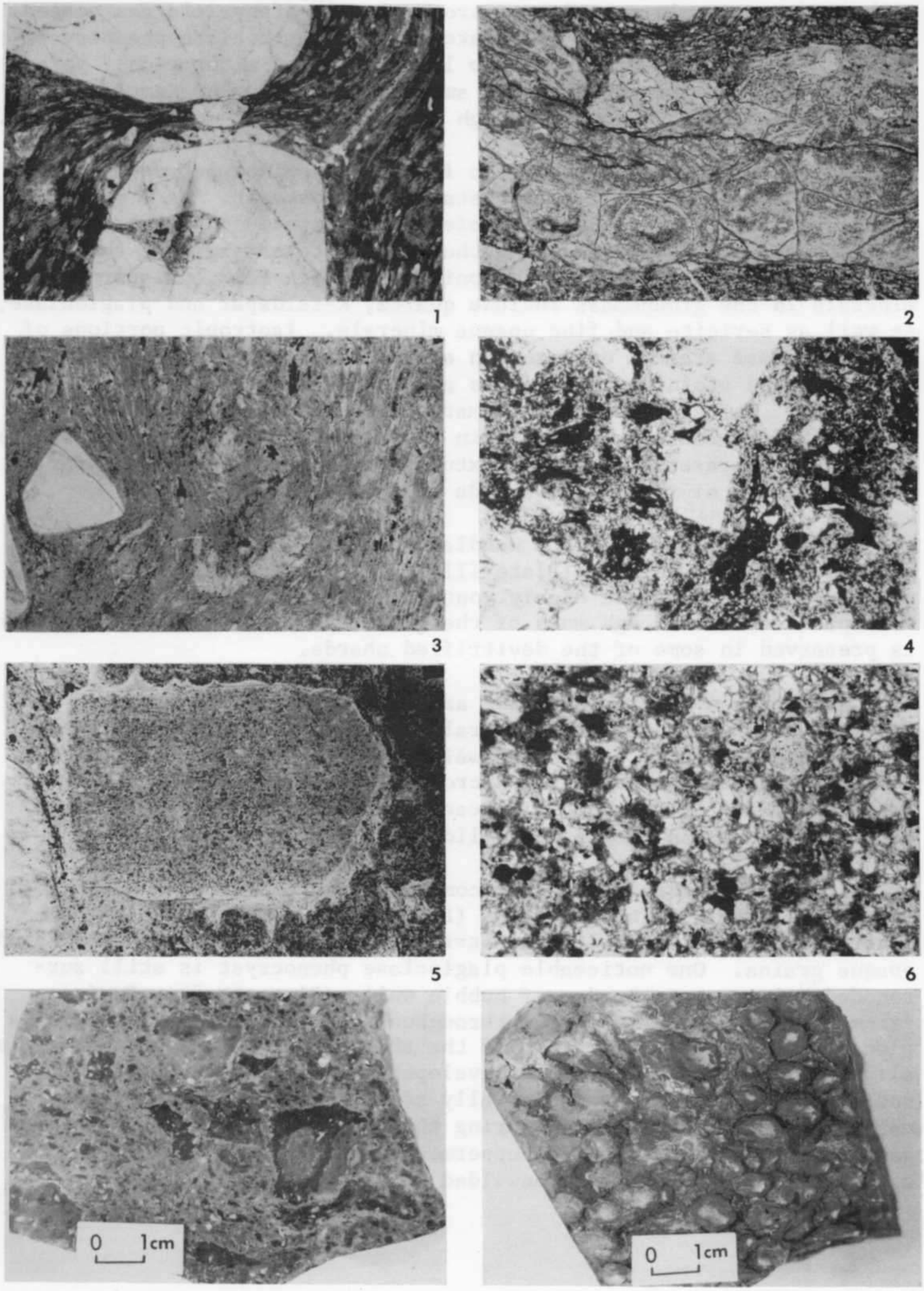


Plate IV - Taylor Formation - Volcanic Rocks

Plate IV.1 Taylor Formation, north of Mt. Orndorf. Felsite with embayed quartz phenocrysts and flow banding. Plate light, 25x. (ES-136)

Plate IV.2 Taylor Formation, north of Mt. Orndorf. Perlitic cracks. Plane light, 25x. (ES-136)

Plate IV.3 Taylor Formation, Mt. Ehrenspeck. Felsite with phenocrysts of quartz and groundmass with welded shard structure. Plane light, 25x. (ES-53)

Plate IV.4 Taylor Formation, Lubbock Ridge. Felsite with quartz and plagioclase phenocrysts and groundmass with unwelded shards and pumice fragments. Plane light, 25x, section stained. (ES-13)

Plate IV.5 Taylor Formation, Lubbock Ridge. Plagioclase phenocryst rimmed by relic bubble walls. Plane light, 25x. (ES-13)

Plate IV.6 Taylor Formation, Taylor Nunatak. Ash-fall tuff with quartz and plagioclase fragments and mottled groundmass. Plane light, 63x. (ES-207)

Plate IV.7 Taylor Formation, north of Mt. Orndorf. Lapilli tuff. Polished slab. (ES-130)

Plate IV.8 Taylor Formation, north of Mt. Orndorf. Accretionary lapilli. (ES-02)

The preceding three samples conclusively prove the existence of ash-flow tuffs in the Taylor Formation. Welded-shard structure is the single most important criterion for their recognition. Ross and Smith (1961) also summarized other characteristics which serve to identify and distinguish ash-flow tuffs, ash-fall tuffs and lavas. Ash-flow tuffs are typically nonbedded and nonsorted, whereas ash-fall tuffs usually exhibit pronounced bedding, and generally contain only material of ash or fine ash size (< 4 mm). Axiolitic structure has apparently been observed only in ash-flow tuff. Numerous foreign rock fragments, though usually amounting to less than 5% of the rock, often accompany tuffs, unlike lavas which contain few or none. Silicic lavas also almost always show flow banding, although this in some cases may be difficult to distinguish from welded shard structure. The discontinuous nature of linear elements may also serve to identify a welded tuff.

The nonbedded, nonsorted nature of the porphyritic felsites of the Taylor Formation, and the common occurrence of volcanic rock fragments in them, strongly suggests that many if not most of these rocks, originated in Peléan-type eruptions of explosive gas and ash, and were deposited as ash-flow tuffs.

Chemical analyses indicate that these porphyritic felsites are rhyolitic in composition (see page 57). Although recent rhyolitic eruptions are in some cases of a Peléan type, often they occur as domes and thick lava flows (e.g., Rose, 1972). Therefore, this latter mechanism should not be completely ruled out for all of the felsites lacking indicative groundmass textures.

A different sort of volcanic rock is represented by several samples from Taylor Nunatak (Plate IV.6). These are composed of many very fine laths of plagioclase and fragments of quartz, plagioclase, and K-feldspar (< 0.125 mm) with long axes vaguely aligned. They are dispersed in a cryptocrystalline groundmass containing finely disseminated opaque minerals and areas which on staining, are seen to be potassium rich. One sample from higher in the same section exhibits a distinct layering with long axes of crystals and fragments, ranging up to 0.8 mm, clearly aligned. All appearances indicate that these rocks originated as ash-fall tuffs.

In all cases, samples from units which in the field appeared to be chert, were found to contain at least some very fine crystal fragments and microcrystalline feldspar in addition to quartz, so that an origin as volcanic ash seems indicated. The occurrences not sampled for petrographic examination are herein called "chert" since it is expected, but not proven, that they are not pure silica and are probably of volcanic origin.

Two examples of a lapilli tuff were found. One from north of Mt. Orndorf contains numerous lapilli, up to about 2 cm, with ragged, flame-like edges (Plate IV.7). The lapilli are light green to dark green, usually somewhat chloritic and float in a light green groundmass without chlorite. Quartz and plagioclase phenocrysts occur in both the lapilli and the groundmass.

Variation of lapilli is greater in a sample from Taylor Nunatak-South. Included are basic types with plagioclase phenocrysts in a dark groundmass of plagioclase microlites and opaque minerals, as well as more silicic varieties colored red and gray with both quartz and feldspar phenocrysts. The form and distribution of the lapilli are similar to that of the sample at Mt. Orndorf with sizes ranging to more than 3 cm. The microcrystalline groundmass contains occasional patches of sparry calcite.

A specimen from high on a ridge crest north of Mt. Orndorf, not in place but probably locally derived, was found to contain accretionary lapilli (Plate IV.8). The accretionary lapilli occur as usually intact, aligned ellipsoids with long axes 5-10 mm in length, and packed so that adjacent ellipsoids are in contact. Compositionally the lapilli are microcrystalline quartz and feldspar, calcite, and minor amounts of opaque minerals, with some small (to 0.1 mm) crystal fragments of quartz and plagioclase arranged concentrically in each ellipsoid. A slight decrease of grain size occurs outward in the ellipsoids and most have an outer shell (5 mm) of slightly darker material. The surrounding matrix is composed of similar material, along with broken crystals of plagioclase and quartz (to 1 cm) and patches of sparry calcite. Also present are forms suggesting devitrified glass shards.

Moore and Peck (1962) in summarizing the occurrence of accretionary lapilli conclude that they are formed by accretion of moist ash in an airborne cloud during a Peléan-type eruption, and for historically observed eruptions, with a few exceptions, they have fallen within 16 km of the volcanic vent onto land or into shallow water with rapid burial, since long immersion would lead to their disintegration.

Sedimentary Rocks

Lower Units

The lowermost metasedimentary rocks in the Shackleton Glacier area occur on the eastern slopes of Mt. Greenlee and Epidote Peak. Wade (1974) named the sequence underlying the first metavolcanic rocks the Greenlee Formation, although similar rocks persist into the lower portion of the overlying Taylor Formation where they are intercalated with basaltic lavas and felsites.

The metasedimentary rocks are fine-grained quartzites and phyllites derived from feldspathic graywackes, quartz siltstones and shales. Some are slightly calcareous. Bedding is even and ranges from 5 cm-1 m. Fine laminations are the only observable sedimentary structures. Metamorphism has recrystallized the pelitic fraction to biotite and in some cases biotite-hornblende-epidote. Foliation parallel to bedding is developed in the metasedimentary rocks on Epidote Peak but is not so distinct at Mt. Greenlee. The grain size and sedimentary structure indicate that these metasediments were deposited in a very low energy environment.

The section continues on the ridges west and northwest of Epidote Peak where there appears to be a mixed sequence of metagraywackes and felsic metavolcanics. Several white, coarsely crystalline marble units are found at the westernmost ends of the ridges. Unfortunately relationships are obscured throughout by considerable cataclasis.

Middle Units

Overlying the massive volcanic rocks at the base of Taylor Nunatak-North is a varied 200 m section of limestone, "chert," coarse clastic sedimentary rocks and volcanic rocks. Lithologies change rapidly across the strata with individual units seldom exceeding 20 m.

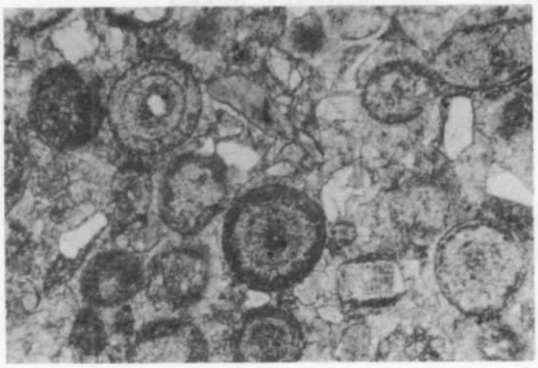
A variety of calcareous sediments are represented. Micrites occur at several levels, often with continuous or discontinuous interbedded "chert". In some of the micrites layering within the beds has been turbated.

Clastic oosparite crops out in two horizons (Plate V.1). Grains are spaced loosely with point contacts and infilled by medium crystalline (0.0625-0.25 mm) sparry calcite. The oolites range up to 0.7 mm in diameter and exhibit concentric structure. A few are broken. Nuclei in some cases are angular grains of quartz or micritic intraclasts. Very fine-grained quartz has selectively replaced some or all of the layers in some of the oolites.

The clastic fraction includes angular to sub-angular quartz and plagioclase grains and fragments of silicic, volcanic rock and "chert". The quartz is often embayed, confirming the volcanic origin of the grains. These rocks are indicative of a shallow, high-energy, marine environment adjacent to a volcanic source area.

One interesting sample contains ragged, volcanic lapilli (up to 0.5 cm) floating in a matrix of sparry calcite (Plate V.2).

PLATE V – TAYLOR FORMATION, SEDIMENTARY ROCKS



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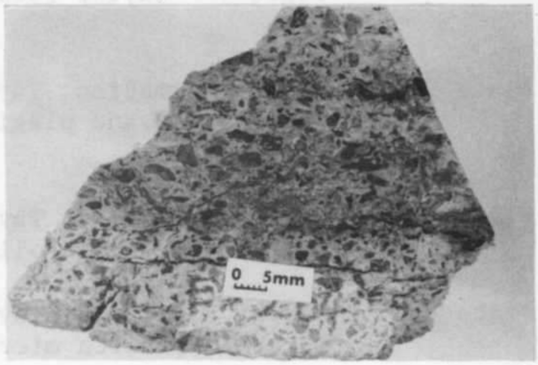

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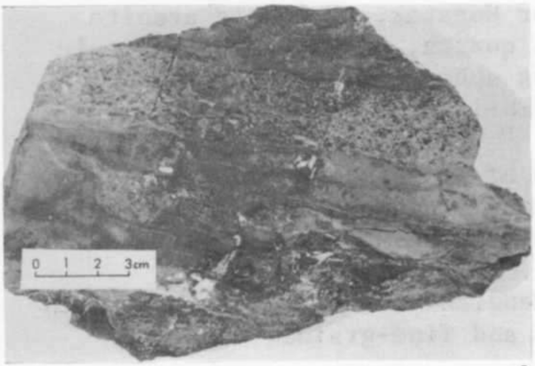

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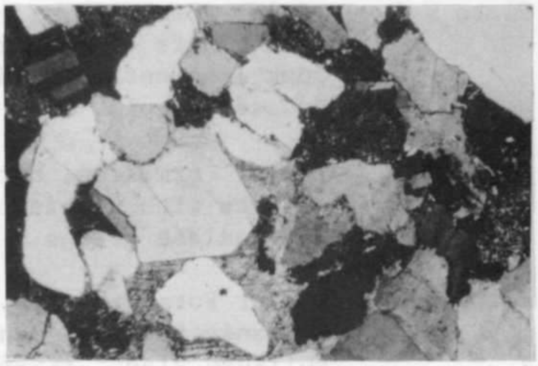

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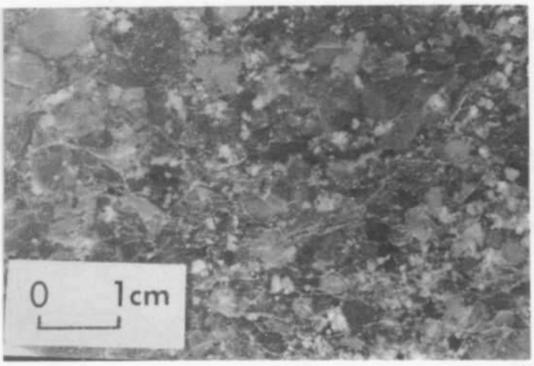

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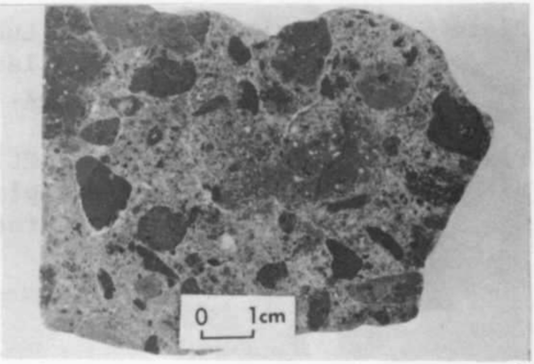

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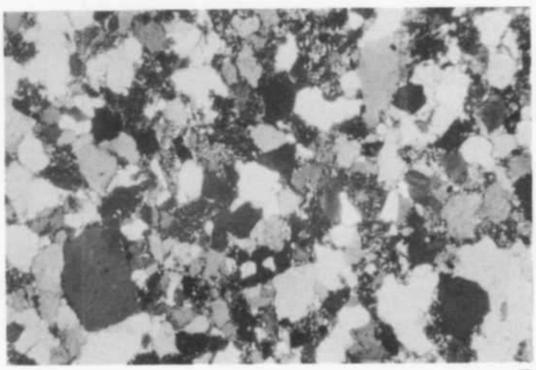

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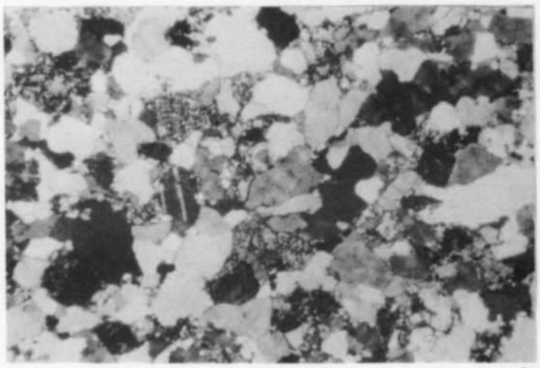

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Plate V - Taylor Formation - Sedimentary Rocks

Plate V.1 Taylor Formation, Taylor Nunatak. Clastic oosparite with quartz and plagioclase clasts. Plane light, 25x. (ES-191)

Plate V.2 Taylor Formation, Taylor Nunatak. Volcanic lapilli in sparry calcite. Polished slab. (ES-204)

Plate V.3 Taylor Formation, Taylor Nunatak. Volcanic arenite interbedded with micrite. Polished slab. (ES-194)

Plate V.4 Taylor Formation, Taylor Nunatak. Volcanic arenite with clasts of euhedral quartz, plagioclase, volcanic rock fragments including spherulitic rock fragment. Crossed nicols, 25x. (ES-182)

Plate V.5 Taylor Formation, north of Mt. Orndorf. Volcanic arenite with volcanic rock fragments and quartz and plagioclase grains. Polished slab. (ES-134)

Plate V.6 Taylor Formation, Mt. Wendland. Laharic breccia, with volcanic rock fragments and fine-grained matrix. Polished slab. (ES-65)

Plate V.7 Taylor Formation, Lubbock Ridge; Subarkose with clasts of quartz, plagioclase and volcanic rock fragments. Crossed nicols, 25x. (ES-20)

Plate V.8 Taylor Formation, Mt. Ehernspeck. Subarkose with clasts of quartz, plagioclase, K-feldspar and volcanic rock fragments. Crossed nicols, 25x. (ES-44)

A 9 m thick fossiliferous breccia containing fragments up to 3 cm in length occurs in this section. Lithologies of the breccia fragments include (1) micritic limestone intraclasts, some of which are pelleted, (2) felsite rock fragments, some porphyritic and some spherulitic, (3) occasional lava fragments with plagioclase microlites and phenocrysts, and (4) euhedral plagioclase and rounded, embayed quartz crystals to 1.5 mm. These are loosely set in a matrix of fine-grained angular quartz and plagioclase, oolites, and finely crystalline spar.

The bimodal size distribution of clasts and matrix, the heterogenous character of the clasts, and the lack of clast fabric suggests that this unit was deposited by a process of mass movement, perhaps as a debris flow which slid from a bank margin into deeper water, in a manner analogous to deposits described by Cook and others (1972).

Small, tube-shaped fossils, the first found in the Taylor Formation, occur in the matrix material. They have been identified as Cloudina? and given an Early Cambrian age (Yochelson and Stump, (1977). These fossils have affinities with the little known Cloudina? borrelloi from the Lower Cambrian of San Juan Province, Argentina (Yochelson and Herrera, 1975), but are distinct from Cloudina of the Nama of South West Africa discussed in Chapter 11 (Germs, 1972b).

Rocks consisting of finely laminated, recrystallized ash and fine-grained quartz, plagioclase, and K-feldspar fragments occur at levels in the section. The fineness of these sediments and their even layering indicates that they accumulated in a quiet water setting, probably below surf zone, without the influence of traction currents. Graded bedding can be seen with the microscope in some layers less than 1 mm thick. This surely represents ash-falls which settled through the water in succeeding waves of fine debris.

In association with these fine-grained ash deposits and with some micrites are coarse-grained volcaniclastic deposits with no fine matrix material between the grains. The sand is composed primarily of subangular and euhedral plagioclase and embayed quartz, appearing to be a concentration of volcanic phenocrysts washed free of their groundmass. Also present are silicic volcanic rock fragments, some containing phenocrysts and some with spherulites (see Plate V.4). One sample contains a considerable number of pumice fragments and fragments containing unwelded shard structure.

These volcaniclastic units are sharply bounded above and below by the finer ash or micrite, whose upper surfaces appear in some cases to have been disturbed during deposition of the overlying coarse-grained sediments (see Plate V.3). The sandstone units are internally massive, but in some cases indistinct layering due to size grading of the clasts is apparent.

The presence of these coarse-grained sedimentary rocks in association with sedimentary rocks deposited under quiet water conditions indicates that they have probably been resedimented. An environment of energetic currents would be needed to concentrate the coarse-grained sand prior to its final deposition in quiet water. The lack of internal bedding and sedimentary structures except for indistinct layering, and the sharp boundaries to the units suggest deposition by some sort of grain-flow process (Sanders, 1975; Stauffer, 1967; Middleton and Hampton, 1973), with the volcaniclastic sediments flowing downslope into an environment of deeper and quiter water.

Upper Units

In the area of outcrops from Nilsen Peak to Lubbock Ridge is the third suite of rocks of the Taylor Formation, characterized by pyroclastic, volcaniclastic and clastic lithologies. Areas in the north appear to be underlain by a facies which, in addition to ash-flow units, includes waterlaid ash, carbonates, and fine to coarse, volcanically-derived clastics, in contrast to the southern facies which contains argillites and clean, cross-bedded, feldspathic quartzites interbedded with ash-flow tuffs.

Interspersed with the porphyritic felsites north of Mt. Orndorf are several hundred meters of dark green to gray, chert-like rock which is interpreted as, at least in part, water-laid ash deposits. Under the microscope much of this material appears gray and cryptocrystalline, with a sprinkling of opaque grains and possibly calcite. The rock is usually bedded (3-50 cm) and contains graded-beds, cross-beds, scours, and laminations. Conglomeratic layers occur rarely as well, with rounded or subangular clasts of silicic volcanic rock similar to the surrounding beds. Portions may be calcareous with thin (5 mm) graded-beds containing coarse calcite.

A 30-40 m carbonate unit crops out in the sequence north of Mt. Orndorf. The lower half is a white, sucrose marble alternating at about 20 cm intervals with 3-5 cm, discontinuous layers of gray-green, sandy shale. The upper half contains uneven, alternating layers with white and maroon, green or cream hues. The white portion is a marble with about 10% angular to sub-angular quartz and lesser amounts of plagioclase spread throughout medium-crystalline spar. The colorful layers contain similar quartz and feldspar, but the most abundant clasts are subangular silicic volcanic rock fragments set in a recrystallized groundmass, in part containing epidote. The ease with which this clastic material filled around broken pieces of marble suggests that this is a preconsolidation texture rather than a metamorphic one, since carbonate rocks flow more readily under elevated pressure-temperature conditions than silicate rocks.

Volcaniclastic conglomerate occurs both north of Mt. Orndorf and near Mt. Wendland. Washed free of finer material, the poorly-sorted medium-sand to pebble size clasts are composed of silicic volcanic rock fragments, some containing spherulites, some with plagioclase microlites, and volcanically derived, embayed quartz, euhedral plagioclase and sometimes orthoclase. The rock fragments are rounded to sub-angular and the crystal grains appear to have suffered little abrasion. Textural features include clast alignment and a crude layering. Transport by strong currents and rapid deposition accounts for the poorly-sorted but unsilty character of these sediments.

South of Mt. Wendland several different types of clastic rocks also occur. Cross-bedded, feldspathic quartzites and phyllites are present, as is a stretched conglomerate, pictured by McGregor (1965), containing boulders of volcanic rock and marble. In addition there is a 100 m sequence of massive poorly sorted conglomerate intermixed with coarse sandstone layers. The larger clasts in the conglomeratic horizons are a variety of silicic volcanic rock fragments, whose maximum size of about 30 cm near the bottom decreases upwards in the section to about 3 cm at the top. Quartz, plagioclase, and orthoclase crystals of volcanic origin are plentiful, and unlike most of the other conglomerates and coarse quartzites, a microcrystalline matrix is abundant.

Interbedded with porphyritic felsites from south of Mt. Kenny to Lubbock Ridge are thin-bedded, dark gray to green argillites and festoon cross-bedded, white to light gray quartzites. The quartzites are medium- to very coarse-grained, subangular to subrounded, and well sorted with bedding ranging from 3-60 cm. Clasts include quartz, plagioclase, sometimes orthoclase, and silicic volcanic rock fragments, with negligible matrix material (Plate V.7, 8) (Table 5).

Table 5. Modal analyses (in percent) of sedimentary samples from the Taylor Formation (300 point counts per sample).

Rock Name	Sample	Quartz	Plag	K-spar	Total R.F.	Calcite	Matrix
Subarkose	ES-20	72	13(total fsp)		7	t	8
Subarkose	ES-44	78	5	6	4	0	7
Lithic Arenite	ES-182	38	20	0	24	6	12
Feldspathic Graywacks	ES-185	33	24	1	16	2	24

Quartz grains are usually sutured at contacts and plagioclase is altered to sericite. Occasional quartz grains preserve embayments, but a small number of rutilated quartz and perthite grains indicate derivation of sediment from a plutonic provenance for the first time in the Taylor Formation. Rock types and sedimentary structures indicate an environment of active fluvial or shallow marine currents, disrupted by intermittent pyroclastic eruptions.

FAIRWEATHER FORMATION

Introduction

The Fairweather Formation is exposed on Mt. Fairweather and the ridge crest extending eastward from it to the Duncan Mountains, on spurs #19 and #20 of the Duncan Mountains, on the ridge west of Mt. Henson, and in the area of Mt. Roth and Mt. Justman (see Appendix A for location maps). McGregor (1965) named the formation and designated Mt. Fairweather and Fairweather Ridge-East as the informal type locality. There the formation is asymmetrically folded with overturning to the south and strike perpendicular to the ridge crest so that a nearly complete cross-section is exposed. In 1974-1975 our field party refined the structure delineated by McGregor and measured the section from the summit of Mt. Fairweather to its fault contact with the Duncan Formation (see Figure 4). A breccia developed during syntectonic intrusion occupies almost a quarter of the section, and whereas the fragments are made up of a single type of metavolcanic rock, one can only speculate on the actual thickness prior to intrusion.

With basic and acidic volcanic rocks, "chert" and marble beds, and coarse clastic and volcaniclastic units, the lithology as well as the succession of the Fairweather Formation is similar to that of the Taylor Formation, 100 km to the northwest. However, the Fairweather Formation has suffered a greater degree of metamorphism, such that when studied petrographically the rocks are seen to preserve precious little of their original volcanic or sedimentary textures. As a result premetamorphic characteristics are best inferred from the outcrops.

Stratigraphic Characteristics

The descriptions which follow are arranged by rock type beginning with units lower in the succession. A measured section is produced in Appendix A--Duncan Mountains.

On Mt. Fairweather low in the section are repeated units of metabasaltic rocks or greenstones which by mid-section are reduced to isolated occurrences. These rocks are more often nonfoliated but foliated varieties are not uncommon. In thin section they are seen to consist of intergrowths of epidote, epidote-actinolite, biotite-actionolite and biotite-chlorite, with plagioclase and sometimes small amounts of quartz. In addition, sphene is abundant in several specimens.

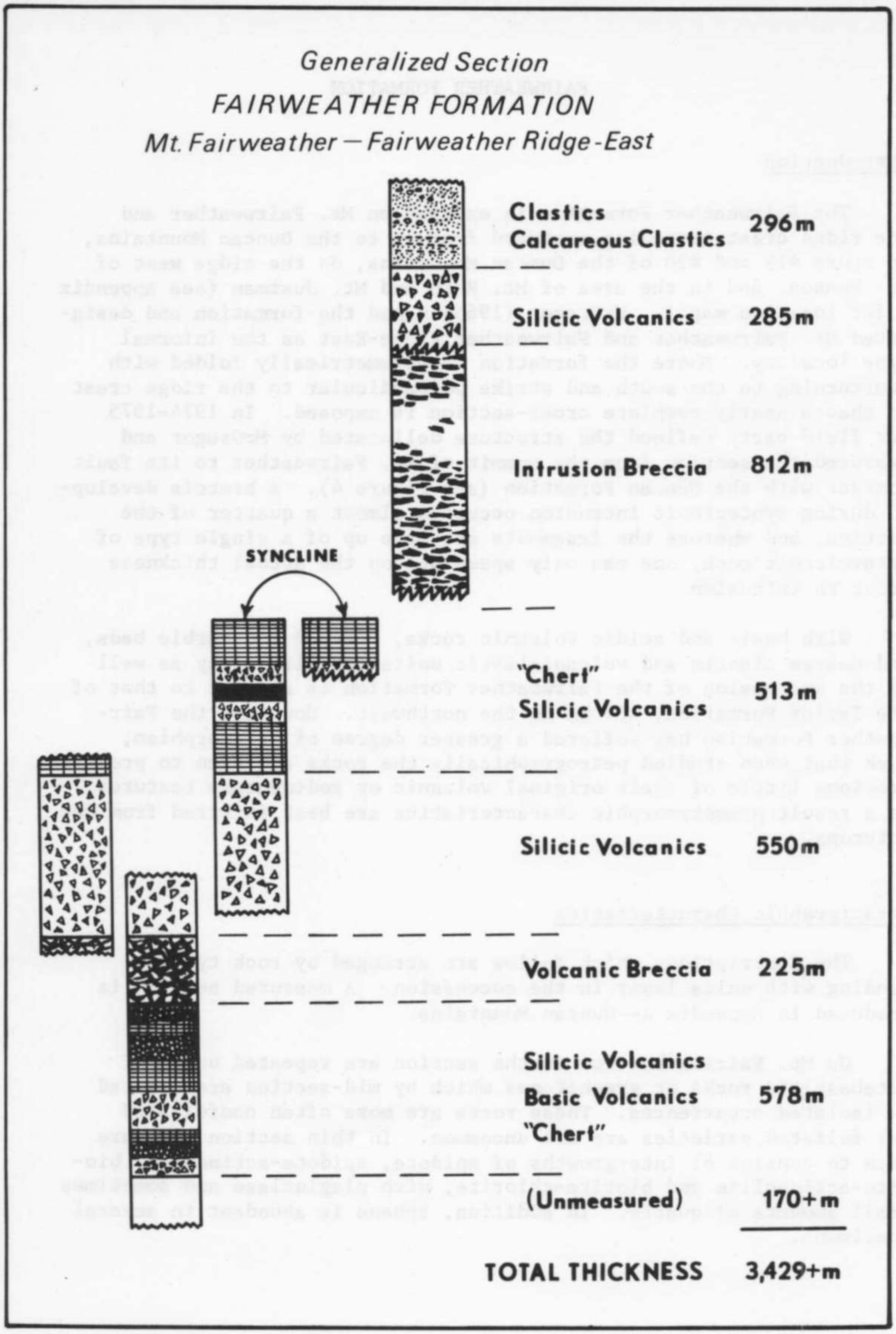

Fig. 4. Generalized section Fairweather Formation, Mt. Fairweather - Fairweather Ridge-East.

Many of the units have suffered considerable post-emplacement alteration. For example, networks of epidote veins sometimes constitute as much as 50% of some samples. That at least some of these rocks originated as extrusive lavas is confirmed by abundant vesicles in certain units (Plate VI.1).

A texture suggesting a basaltic breccia or agglomerate is found near the summit of Mt. Fairweather. This rock contains plagioclase-rich fragments to 20 cm in length in loose contact with one another, presumed originally to have been volcanic breccia fragments. They are surrounded by material rich in epidote with some actinolite and calcite. Except for the aforementioned no other extrusive characteristics (e.g., pillow-structure or layering) were observed, although subsequent alteration and deformation may account for the paucity of recognizable flow features.

Blastoporphyritic metafelsites ranging from a few meters to 550 m in thickness are found in all but the highest levels of the section. Units are massive or foliated and come in a variety of colors, including black, gray, brown, tan, light green, and orange, with lighter shades being more common. Phenocrysts of quartz and plagioclase and/or K-feldspar serve to identify these rocks. In some of the better preserved examples embayments can be seen in the quartz but often the crystals have been stretched and polygonized. Plagioclase crystals, which are often euhedral and twinned, usually have been sericitized or saussuritized. According to the unit, the K-feldspars may be either microcline or orthoclase. Besides the grid twinning of the microcline, Baveno twins sometimes occur.

Abundant angular crystal fragments of quartz and feldspar are the only megacrysts in some units but more often they accompany larger euhedral or embayed feldspar and quartz crystals. The phenocrysts and crystal fragments are usually wholly disordered and suspended in the groundmass, though in a few samples a long-axis alignment and crude layering can be detected.

In all cases the groundmass is at an advanced stage of recrystallization. Fine to medium-grained hornfels or foliated textures completely replace any volcanic textures which may have been formed during emplacement. The mineralogy of the groundmass includes quartz and plagioclase and/or K-feldspar, with chlorite and/or biotite usually intergrown around the framework silicates. Muscovite sometimes accompanies these micas, and fine-grained amphibole occasionally occurs. Extremely fine opaque grains are sprinkled throughout many of the samples.

Except for metamorphic differences between the felsites of the Fairweather and Taylor Formations, strong similarities exist in

PLATE VI — FAIRWEATHER FORMATION

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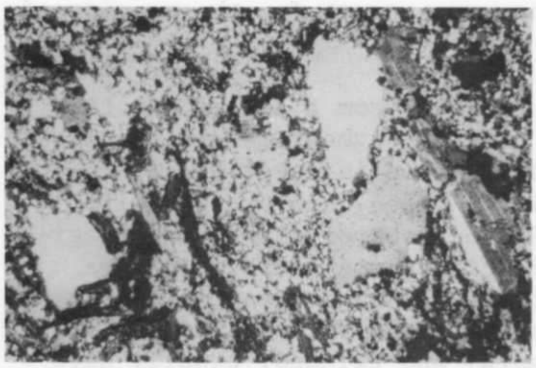

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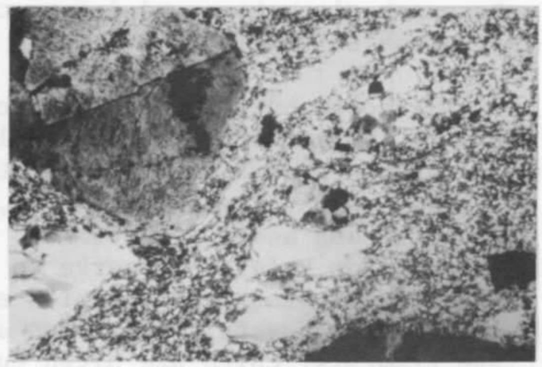

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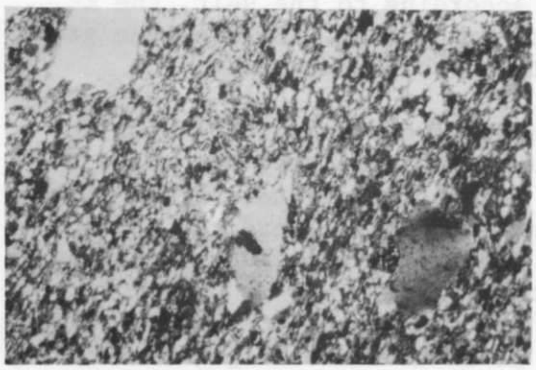

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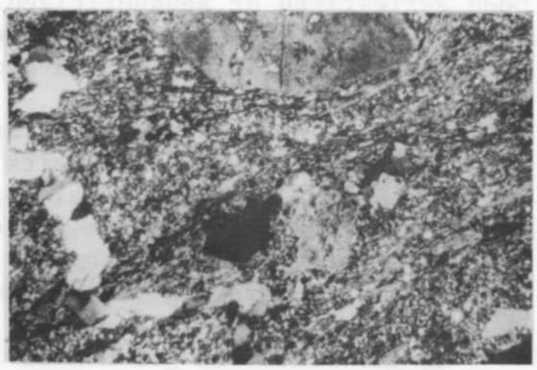

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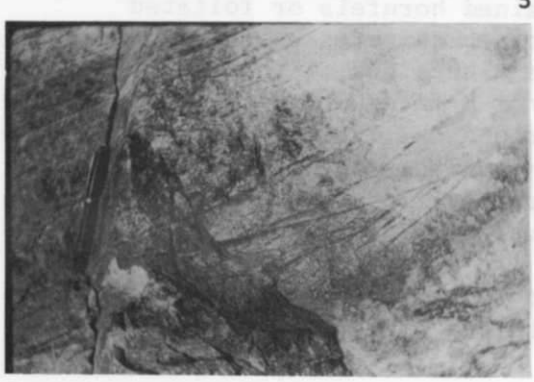

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Plate VI - Fairweather Formation

Plate VI.1 Fairweather Formation, Mt. Fairweather. Amygdaloidal basalt. Side view.

Plate VI.2 Fairweather Formation, Fairweather Ridge-East. Metabasalt with biotite, actinolite, quartz and plagioclase. Plane light, 63x. (ATS)

Plate VI.3 Fairweather Formation, Fairweather Ridge-East. Metafelsite with phenocrysts of quartz and plagioclase and groundmass of quartz, plagioclase, actinolite and minor biotite and chlorite. Crossed nicols, 25x. (ATQ)

Plate VI.4 Fairweather Formation, Fairweather Ridge-East. Metafelsite with phenocrysts of orthoclase (Baveno twin) and quartz and groundmass of quartz, orthoclase and muscovite. Crossed nocols, 25x. (AVG)

Plate VI.5 Fairweather Formation, Fairweather Ridge-East. Metafelsite with phenocrysts of quartz and groundmass of quartz, plagioclase, biotite, chlorite and minor calcite. Crossed nicols, 25x. (ATJ)

Plate VI.6 Fairweather Formation, Fairweather Ridge-East. Metafelsite with phenocrysts of quartz and orthoclase and groundmass of quartz, muscovite, biotite and K-feldspar. Crossed nicols, 25x. (ASJ)

Plate VI.7 Fairweather Formation, Fairweather Ridge-East. Cross-bedding in quartzite.

Plate VI.8 Fairweather Formation, Spur 20, Duncan Mountains. Breccia.

their field aspects, porphyritic textures, mineralogy and chemical compositions. It seems likely therefore that some and perhaps most of the metafelsites of the Fairweather Formation originated by processes similar to those operating during the accumulation of the Taylor Formation, namely Peléan-type explosive eruptions.

Rocks which in the field have the appearance of chert are common in the lower and middle portion of the Fairweather Formation. They may be massive or finely laminated and come in a variety of colors including black, dark brown and green, buff and white. In addition to extremely fine-grained quartz these rocks can be seen in thin section to contain finely disseminated opaque grains, often biotite or hornblende or both, and a low birefringent mineral, presumably feldspar. This assemblage indicates that the original rock was not pure silica when formed. The field characteristics of these rocks are similar to some of the chert-like rocks of the Taylor Formation which have been interpreted as water-laid ash deposits, but the Fairweather occurrences apparently lack the fine crystal fragments found in many of the Taylor samples, so caution must be exercised in using such an interpretation for the origin of these "cherts." Chert sedimentation is recognized as a common association in volcanic island terrains (Kanmera, 1974).

The uppermost portion of the Fairweather Formation contains a suite of clastic and calcareous metasedimentary rocks which in the exposures on the crest of Fairweather Ridge-East are badly sheared and attenuated. However at the base of the ridge 1 km to the northwest the original sedimentary characteristics are well preserved. Outcrops on Spurs #20 and #1 contain similar lithologies and appear to belong to approximately the same stratigraphic level as the metasedimentary rocks at Fairweather Ridge-East. Rock types include metaconglomerate and breccia, cross-bedded pelitic and calcareous quartzite, fine-grained, calcareous hornfels and impure marble.

Much of the northeast side of Spur #20 is underlain by schists interbedded with metaconglomerate or breccia whose clasts and matrix are composed of fine-grained schists with varying amounts of quartz and muscovite. There is continuity of foliation between the clasts and matrix and overprinting by large porphyroblasts of biotite which spot the rock. Some of the clasts contain relic quartz phenocrysts confirming their volcanic provenance while in most the uniform, fine-grain size and composition of the clasts suggest derivation from deposits of silicic ash or, alternatively, very fine-grained quartz arenites.

In most outcrops on Spur #20 the clasts are highly attenuated due to deformation; however, on the crest of the main ridge there is a breccia with its original form well preserved. The fragments which reach 50 cm length are angular and poorly sorted, although average size varies considerably between sub-units. This particular occurrence, and by inference the stretched conglomerates or breccias, originated as rocks slides or slumps which traveled only short distances and prevented rounding even of the largest fragments (see Plate VI.8).

On the southwest side of Spur #20, stratigraphically below the breccias is about 30 m of white, cross-bedded pelitic quartzite followed by 10 m of unevenly layered quartzite and calcareous quartzite. These are overlain to the ridge crest by thin-bedded, impure marbles which are somewhat laminated.

At the base of Fairweather Ridge-East is a mixed clastic and calcareous clastic sequence in which bedding characteristics are well preserved. Conglomerates and breccias, while not as extensive as on Spur #20, occur repeatedly in the section. Light-gray, trough cross-bedded quartzites with beds greater than 1 m are conspicuous. Originally calcareous portions of the section have reacted to produce colorful areas containing tremolite-actinolite, epidote, biotite, garnet, sphene and sometimes calcite. The normal relationship is one of alternating uneven bands containing greater and lesser concentration of calc-silicate minerals. Ellipsoidal concentrations also occur, sometimes in units with cross-bedding. Lastly, one 30 cm bed of pink, pure marble was observed.

The cross-bedded quartzites are indicative of an environment of active currents and the metamorphosed calcareous intercalations suggest that it was shallow marine. The conglomeratic layers were probably derived from nearby sources.

Rock types on Spur #1 are similar to those on Spur #20, and include conglomerate, cross-bedded micaceous quartzite and schist. A distinguishing feature is the presence of occasional orange marble clasts within the conglomerates.

Cross-bedded quartzites which proved to be most useful in determining stratigraphic tops also occur in isolated exposures in the lower and middle portion of the Fairweather Formation. In addition, thin marble beds are a rare, but repeated rock type throughout the section.

CHEMICAL ANALYSES AND SUMMARY OF THE VOLCANIC FORMATIONS

Chemical analyses were run on samples of metavolcanic rocks from the Taylor, Fairweather, and Wyatt Formations in order to classify them and to see if any chemical characteristics exist to distinguish between them. Samples were chosen to represent the range of petrographic types including basalts, silicic lavas, and felsites.

The least altered samples were chosen, but plagioclase has been sericitized to some degree in most samples, and rather severely in a few. Rocks of the Fairweather Formation have all probably been metamorphosed to the hornblende hornfels facies and the Wyatt Formation to the albite-epidote hornfels facies. Although rocks of the Taylor Formation are the least metamorphosed, nevertheless incipient biotite has grown in many samples. In addition Fe_2O_3 exceeds FeO in many of the chemical analyses indicating that the rocks have been oxidized. Because of the alteration, the possibility must be considered that these rocks were affected by metasomatic transport of the more mobile elements.

Classification

The rocks were classified according to the system of Irvine and Baragar (1971). All of the silicic samples are subalkaline according to Figure 5. Plotting the subalkaline rocks on an AFM diagram shows them to belong to the calc-alkaline series (Figure 6). Further subdivision is made using a plot of normative color index versus normative plagioclase (Figure 7). All but two of the samples fall in the rhyolite field. The other two (ASL, a dacite, and ES-307, an andesite) reflect compositions in which CaO exceeds Na_2O and have normative color indices comparable to the rhyolites.

The classification of the basalts is not so clear cut. The two samples from the Taylor Formation (ES-86, ES-99) fall close to the divide between the alkaline and subalkaline fields, whereas the one from the Fairweather Formation (ARM) is clearly in the alkaline field. All of the samples contain less CaO than is normally found in the common basalts. The Al_2O_3 is higher than that ordinarily occurring in either tholeiitic or alkali-olivine basalts, and is at a level usually associated with high-alumina basalt. In addition the Na_2O content is higher and the TiO_2 content lower than that found in alkali-olivine basalt. For ES-99 the Na_2O is higher than that of representative tholeiitic or high alumina basalt as well.

Because of the discrepancies mentioned above it is concluded that these basaltic rocks cannot satisfactorily be assigned to any of the basaltic groups and probably that these rocks have been

Table 7.1 Chemical analyses and CIPW norms of volcanic rocks of the Taylor, Wyatt, and Fairweather Formations, Antarctica

	ES-38	ES-42	ES-45	ES-48	ES-52	ES 193	ES-86	ES-99	ES-234	ES-240	ES-307	ART	ASL	ATB	ARM
SiO_2	76.09	77.21	70.25	77.43	77.31	72.00	49.62	51.38	70.28	66.96	65.20	75.01	71.26	74.20	45.88
TiO_2							0.64	0.51							0.60
Al_2O_3	12.57	12.41	16.51	11.38	12.79	12.19	17.10	17.86	14.12	14.30	15.22	12.19	13.12	12.89	17.95
Fe_2O_3	0.77	0.10	1.80	1.60	1.16	3.90	4.90	4.07	1.54	3.62	2.67	1.87	2.32	1.83	8.51
FeO	0.19	0.31	0.24	0.19	0.76	0.90	5.94	5.15	2.11	2.21	3.42	0.80	1.09	0.28	3.68
MnO							0.19	0.12							0.60
MgO	0.13	0.13	0.28	0.07	1.69	0.43	5.48	4.61	1.06	1.99	2.26	0.86	0.87	0.37	4.20
CaO	0.74	1.01	1.04	0.11	0.13	0.88	8.55	7.31	1.58	1.39	3.04	1.26	1.36	0.65	10.82
Na_2O	2.90	2.51	4.04	0.46	0.96	4.85	3.74	5.32	2.87	2.40	2.45	3.62	1.00	1.40	3.55
K_2O	5.23	5.64	5.46	8.94	3.79	2.34	0.72	0.45	2.80	2.90	2.82	3.22	7.00	6.79	1.95
CO_2	0.03	0.28	0.04	<0.02	0.03	0.08	0.07	0.38	0.85	0.84	0.13	0.07	0.13	0.12	0.17
comb.H_2O	0.53	0.45	0.61	0.24	1.72	1.25	2.42	2.22	1.80	2.37	1.33	0.61	1.23	0.74	1.75
Total	99.18	100.05	100.27	100.42	100.34	98.82	99.37	99.38	99.01	98.98	98.54	99.51	99.38	99.27	99.56
q	35.75	37.05	22.45	38.25	52.65	31.3	0.2		38.2	37.3	29.25	36.4	34.15	37.55	
c	1.6	1.0	3.4	0.9	8.0	1.0	2.6	2.7	6.8	8.0	6.3	1.8	3.2	3.2	
or	32.0	34.0	32.0	54.5	23.5	14.5	4.5	3.0	17.0	18.0	17.5	19.5	43.5	41.5	12.0
ab	27.0	23.0	36.0	4.5	9.0	45.5	35.0	48.5	27.0	22.5	23.0	33.5	9.5	13.0	22.13
ne															17.13
wo	0.3		0.3												0.7
en				0.1		0.4						1.2		0.5	
di wo	0.4		0.8	0.1		0.8	8.7	6.8	0.5	0.4	3.0	1.2	1.2	0.5	13.2
di en	0.4		0.8	0.1		0.8	6.5	4.98	0.28	0.38	1.92	1.2	1.16	0.5	12.2
di fs							2.2	1.82	0.22	0.02	1.08		0.04		0.3
hy en		0.4					9.1	3.3	2.72	5.42	4.68		1.44		
hy fs		0.3					3.1	1.2	2.08	0.58	2.62		0.06		
ol fo								3.39							
ol fa								1.26							
mt	0.6	0.15	0.6	0.6		2.4	5.25	4.35	1.65	4.5	2.85	1.8	2.55	0.6	9.45
hm	0.2		0.9	0.8		1.2						0.1			
il							1.0	0.6							0.2
cc		0.6				0.2	0.2	1.0	2.2	2.2	0.4	0.2	0.4	0.4	0.4

Table 7.1 (continued (Explanation of Samples

ES-38	Rhyolite,	Taylor Formation,	Mt. Ehrenspeck
ES-42	Rhyolite,	Taylor Formation,	Mt. Ehrenspeck
ES-45	Rhyolite,	Taylor Formation,	Mt. Ehrenspeck
ES-48	Rhyolite,	Taylor Formation,	Mt. Ehrenspeck
ES-52	Rhyolite,	Taylor Formation,	Mt. Ehrenspeck
ES-193	Rhyolite,	Taylor Formation,	Taylor Nunatak
ES-86	Basalt,	Taylor Formation,	Mt. Greenlee
ES-99	Basalt,	Taylor Formation,	Mt. Greenlee
ES-234	Rhyolite,	Wyatt Formation,	Blackwall Glacier
ES-240	Rhyolite,	Wyatt Formation,	Blackwall Glacier
ES-307	Andesite,	Wyatt Formation,	Moraine Canyon
ART	Rhyolite	Fairweather Formation,	Mt. Fairweather
ASL	Dacite,	Fairweather Formation,	Mt. Fairweather
ATB	Rhyolite,	Fairweather Formation,	Fairweather Ridge-East
ARM	Basalt,	Fairweather Formation	Mt. Fairweather

Analyses performed by Andrew S. McCreath & Son, Inc.

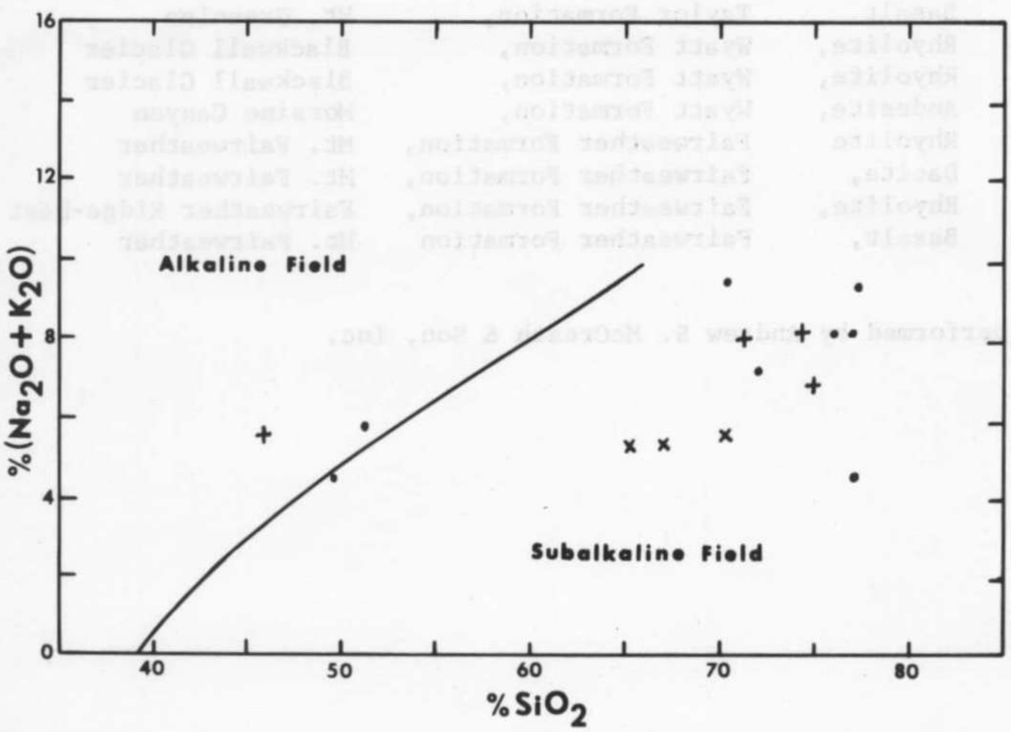


Figure 5. Alkalies-silica plot

Sample designations:
. Taylor Formation
+ Fairweather Formation
x Wyatt Formation

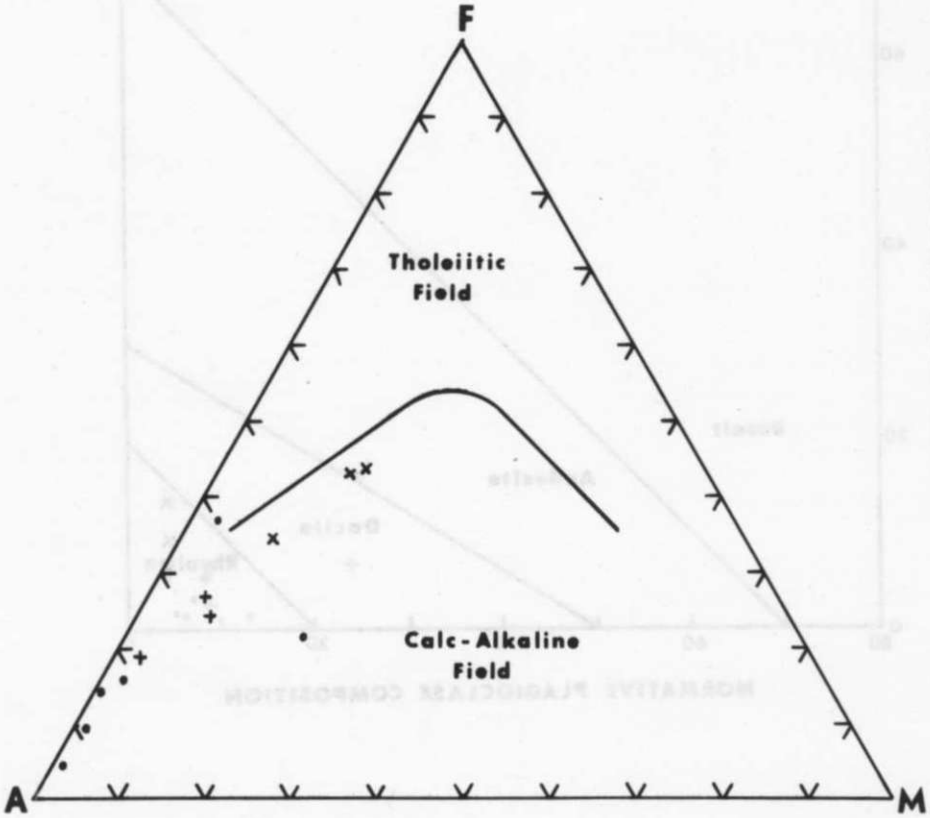


Figure 6. AFM plot

Sample Designations:
. Taylor Formation
+ Fairweather Formation
x Wyatt Formation

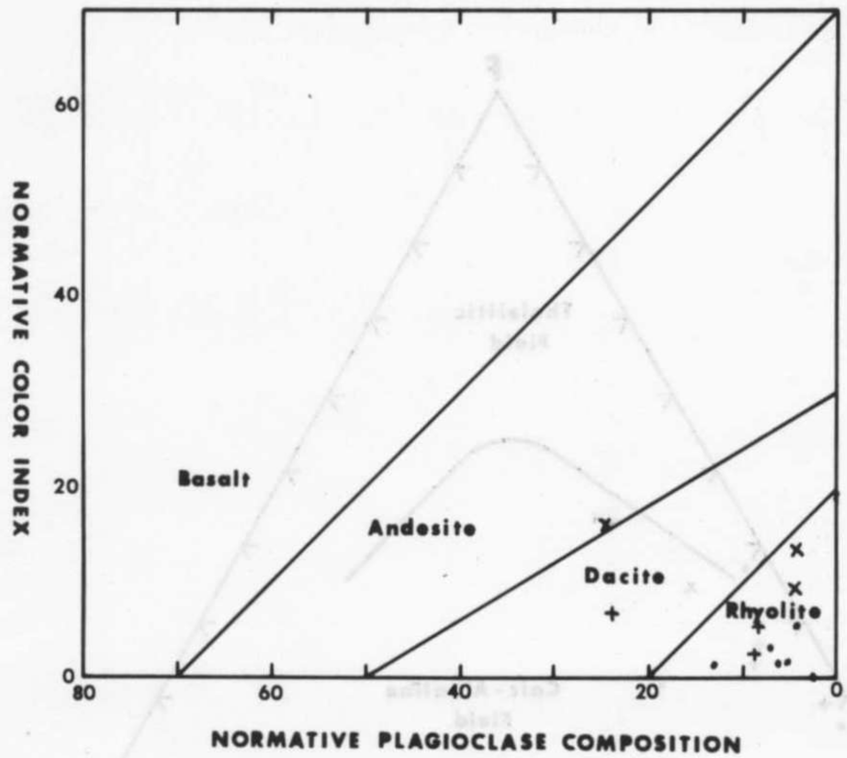


Figure 7. Plot of normative color index versus normative plagioclase composition

Sample Designations:
. Taylor Formation
+ Fairweather Formation
x Wyatt Formation

effected by the removal of CaO and the addition of Na_2O subsequent to their extrusion.

The preponderance of rhyolites is anomalous when compared to calc-alkaline volcanic provinces of Mesozoic or Tertiary age elsewhere in the world, where andesite with lesser dacite and rhyolite is the normal association. The interpretation of the rhyolites originating as ash-flow tuffs (pp. 31-48) is consistent with eruptive processes in other calc-alkaline volcanic arcs, even though the compositions are on the whole much more silicic in these rocks. It is possible that metasomatic processes have served to silicify these rocks, or alternatively that this suite of rocks has been predominantly rhyloitic since its eruption. The data are insufficient to properly evaluate the former possibility.

Comparison of Formations

The Wyatt Formation appears to have distinctive characteristics from the silicic volcanic rocks of the Fairweather and Taylor Formations. On the SiO_2 versus $Na_2O + K_2O$ plot the Wyatt samples can be seen to fall in a group separate from those of the other formations (Figure 5). It may also be noted that the Wyatt samples contain 6-8% normative corundum by contrast to the 3% or less for samples of the Fairweather and Taylor Formations, or that the Wyatt are more aluminous and less silicic than the volcanic rocks of the other two formations.

These chemical characteristics, coupled with the difference in ages between the Wyatt and the Taylor Formations (late Precambrian --isotopic dates versus Early Cambrian--fossils), and with the observation that the Wyatt Formation contains phenocrysts of biotite whereas the Taylor and Fairweather Formations do not, strongly suggests that the origin of the Wyatt Formation was distinct from that of the Fairweather and Taylor Formations.

Summary of the Volcanic Formations

The following summarizes the interpreted history of the volcanic basement formations in the Queen Maud Mountains. Reference may be made to the figures pp. 73-102. Calc-alkaline volcanic events commenced in the late Precambrian with porphyries of the Wyatt Formation outpouring onto the turbidite deposits of the LaGorce Formation. Hypabyssal intrusion of the porphyry also occurred at places. The Wyatt Formation appears to have been emplaced on the craton side of the axis of the later batholithic intrusions of the Ross Orogeny. Across this axis and to the far side of it from the craton, the Taylor and Fairweather Formations were deposited during a period which includes the Early Cambrian.

Basalts were the first extrusions of the Taylor Formation and they occur low in the Fairweather Formation. The basic lavas were succeeded by rhyolites which were erupted as pyroclastic debris that produced an arc of volcanic islands with an associated sedimentary complex that included carbonate banks, oolite shoals and coarse volcaniclastic accumulations deposited by strong current action and sedimentary gravity flows. Erosion of portions of the island arc led to the formation of relatively mature, cross-bedded quartzites throughout the region and coarser conglomeratic deposits locally. These were followed by the development of widespread carbonate banks which are the final deposits recorded in the area.

STRUCTURE AND METAMORPHISM

Introduction

Since the areas visited are widely separated by batholithic intrusions and ice cover, structural continuity is lacking except locally. Also, in most cases in the areas that were examined the structure is relatively simple, especially when compared to the Franke Migmatite (Burgener, 1975) which is outside the scope of this study. Observations of metamorphic effects were made during the course of petrographic studies aimed primarily at determining original volcanic and sedimentary characteristics of the Ross Supergroup.

This chapter will outline the pertinent structural and metamorphic features in each of the several areas studied.

Canyon-Ramsey Glacier Area

Rocks of the Goldie Formation, exposed in the Canyon and Ramsey Glaciers area are invariably steeply dipping. Cleavage is developed in some of the more pelitic parts, usually at a very low angle to bedding. One isoclinal anticline was observed near Gray Peak, west of Canyon Glacier, but otherwise no large scale folding was found. This does not necessarily mean that folding is not present elsewhere, for outcrops are limited, and in fact folding would be expected as the mechanism for producing vertical bedding in an orogenic terrain such as this. Farther to the north, in the Beardmore and Nimrod Glaciers area, isoclinal folding seems to be the rule in the Goldie Formation (Gunner, 1971a; Gunn and Walcott, 1962; Laird and others, 1971).

Biotite and chlorite have grown in the shales and graywackes of the Goldie Formation where examined. Both foliated and hornfels textures are found. Actinolite was found in one calcareous graywacke from Gray Peak.

Shackleton Glacier Area

Beds of the Taylor and Greenlee Formations dip steeply away from Shackleton Glacier on both sides, with less steep dips on Taylor Nunatak. The glacier apparently has removed the crest of a considerable anticline whose core is highly mobilized gneiss, the Franke Migmatite of Burgener (1975), exposed on Mt. Speed, Mt. Wasko, Mt. Franke and Gemeni Nunataks.

At glacier level on Epidote Peak and on the northern end of Mt. Greenlee a zone of highly disordered gneiss, permeated by granitic and pegmatite veins, sharply contacts and cross-cuts at a low angle the Greenlee Formation, which is only slightly metamorphosed and has minimal internal deformation. The evidence suggests an infrastructure-superstructure relationship similar to that in the classic terrain of East Greenland (Wegmann, 1935; Haller, 1955), but not nearly so extensively developed.

Both limbs of the presumed anticline are less deformed toward the core than on their extremities. On the eastern side a sub-vertical zone of mylonite occurs in porphyritic rhyolites of the Taylor Formation adjacent to Massam Glacier between Nilsen Peak and Mt. Orndorf (McGregor, 1965) (Plate VII.1, 2, 3). Augen structure is well developed around the phenocrysts, producing a pronounced lineation nearly down dip on the foliation. The mylonite makes a transition into zones of highly and incipiently foliated rock on the eastern Cathedral Peaks and Lubbock Ridge. This suggests that vertical displacement was concentrated in a highly strained zone at the northern end of this limb of the anticline and that movement was more diffuse toward the south. Franke Migmatites are exposed on Longhorn Spurs across Massam Glacier from the mylonites.

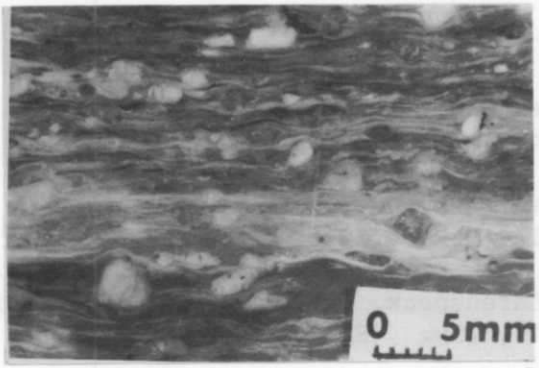
On the western limb of the anitcline the relatively undeformed beds on the lower slopes of Mt. Greenlee and Epidote Peak become increasingly sheared toward the west on the upper ridge crests of Epidote Peak and the unnamed peak north of it, until they disappear beneath snow cover. Much of the rock is massive and silicic, and has been fractured such that the rock weathers into small irregular chips. Metamorphic grade also appears to increase toward the west. A coarsely crystalline white marble unit near the western end of the ridge from Epidote Peak is pinched from 100 m to 10 m in a distance of 100 m. Thin (2-4 cm) silicic beds in the marble are ripped apart and disharmonically folded.

Plutons of post-tectonic Speed Granite engulf portions of both the infrastructure and superstructure of the area. The Taylor Formation is truncated to the east of the Cathedral Peaks and Lubbock Ridge and at glacier level, by cross cutting intrusions. Another pluton occurs at Mt. Speed where it arose in the Franke Migmatite.

Duncan Mountains and Mt. Hensen

Adjacent to the Ross Ice Shelf from Barrett to Scott Glaciers, a group of isolated metamorphic rocks occurs along the exterior margin of the Queen Maud Batholith at Mt. Justman and Mt. Roth, Mt. Henson, the Duncan Mountains, the Herbert Range, and O'Brian Peak. All occurrences are highly deformed and in most cases metamorphosed to the amphibolite or hornblende hornfels facies. The rocks at Mt. Justman and Mt. Roth were incompletely examined so the structural

PLATE VII — STRUCTURE



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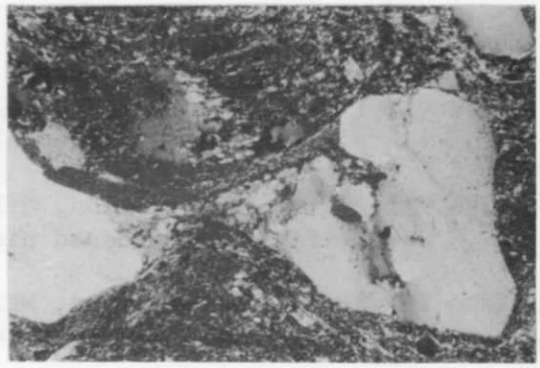

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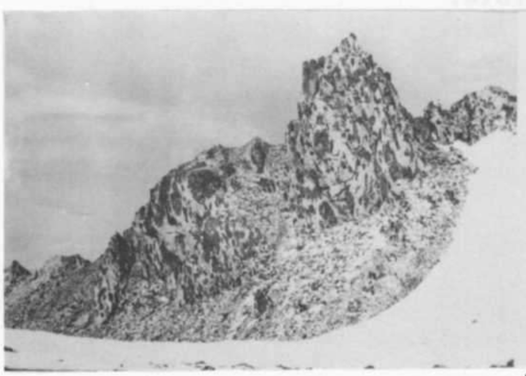

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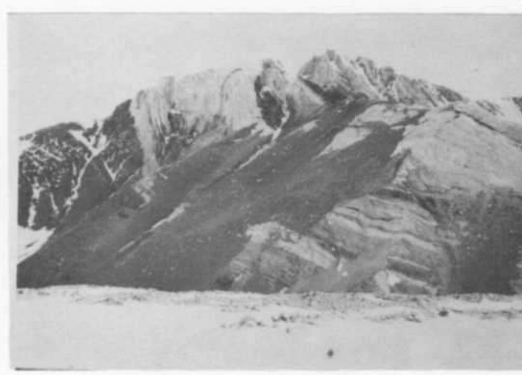

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Plate VII - Structure

Plate VII.1 Taylor Formation, Nilsen Peak. Mylonite in porphyritic felsite. Polished slab.

Plate VII.2 Taylor Formation, Nilsen Peak. Mylonite in porphyritic felsite. Crossed nicols, 25x. (ES-67)

Plate VII.3 Taylor Formation, Mt. Ehrenspeck. Mylonite in porphyritic felsite. Crossed nicols, 25x. (ES-47)

Plate VII.4 Duncan Formation, Duncan Mountains. Asymmetric folds.

Plate VII.5 Fairweather Formation, Fairweather Ridge-East. Intrusion breccia.

Plate VII.6 Fairweather Formation, Fairweather Ridge-East. Intrusion breccia. Detail of Plate VII.5.

Plate VII.7 Mt. Henson, south face. Fault contact between Henson Marble and Duncan Formation.

Plate VII.8 Mt. Henson, north face. Fault contact between Duncan Formation and Henson Marble.

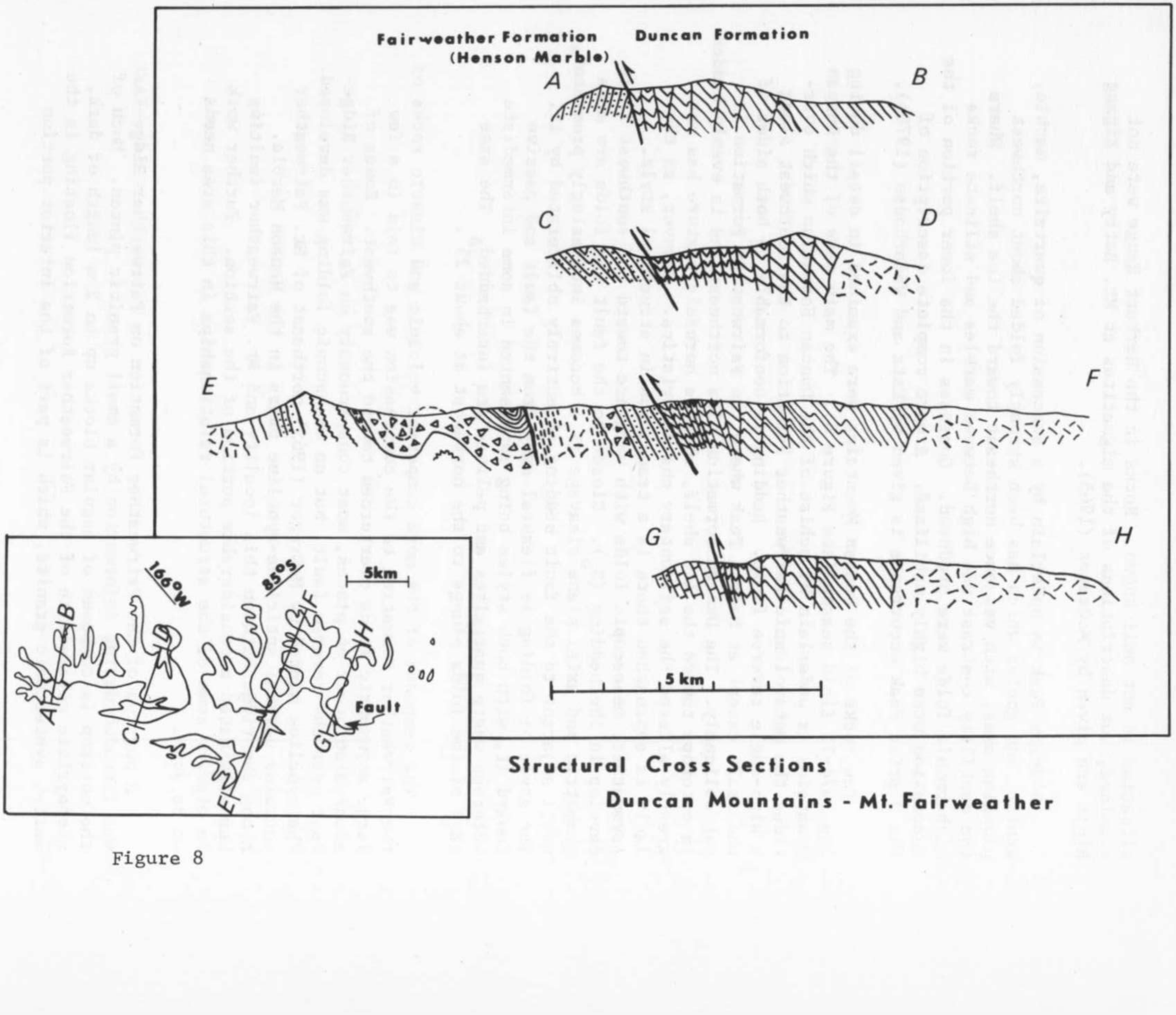
situation is not well known. Rocks in the Herbert Range were not examined, but descriptions of the migmatites at Mt. Betty and Zigzag Bluff are given by McGregor (1965).

O'Brian Peak is underlain by a succession of quartzite, marble, schist, and gneiss which has been strongly folded about northwest plunging axes, with vergence northeast toward the ice shelf. Where the ductility contrast was high between marbles and silicate rocks disharmonic folds were produced. Gneisses in the lower portion of the succession were highly mobilized. A more complete description of the O'Brian Peak occurrence is given by Katz and Waterhouse (1970b).

The rocks of the Duncan Mountains were examined in detail during the 1974-75 field season (see Figure 8). The main mass of the Duncan Mountains is underlain by schist of the Duncan Formation which over-rides the metavolcanic Fairweather Formation to the southwest along a high-angle reverse fault. Bedding is conformable on both sides of the fault except at Morris Peak where the Fairweather Formation is cut obliquely. The Duncan Formation dips northeast and is evenly bedded in outcrops toward the ice shelf, where a hornfels texture has not greatly altered the sedimentary characteristics. However, as the fault is approached there is a transition in structural style. Asymmetric, mesoscopic folds with vergence toward the southwest first develop in the bedding (S_o). Closer to the fault the folds are more symmetric and axial plane cleavage (S_1) becomes increasingly prominent, until adjacent to the fault bedding is entirely obliterated by it. The style of folding is flexural away from the fault and passive toward it, with both styles being represented in some intermediate outcrops where quartzites and pelites are interbedded. The axes (B_1) of the folds plunge to the northwest at about 25^o.

The response of the more competent volcanic and clastic rocks of the Fairweather Formation to the compression was to fold in a few large asymmetric folds overturned toward the southwest. Zones of shear also occur at places, most conspicuously on Fairweather Ridge-East near the reverse fault, but no mesoscopic folding was developed. The syncline mapped by McGregor (1965) northeast of Mt. Fairweather contains several anticline-syncline pairs in the Henson Marble. Along the ridge between this locality and Mt. Fairweather faulting has eliminated a considerable portion of the section. Further work to clarify some of the structural relationships in this area needs to be done.

A portion of the Fairweather Formation on Fairweather Ridge-East was intruded during deformation by a small granitic pluton. Much of the outcrop is composed of angular blocks up to 2 m length of dark, porphyritic metafelsite of the Fairweather Formation floating in the white, syntectonic granite, which is part of the interior portion



Structural Cross Sections
Duncan Mountains - Mt. Fairweather

Figure 8

contains no blocks. The blocks are all aligned vertically in the plane of bedding of the surrounding Fairweather Formation, indicating that the magma was intruded with convergent flow (Balk, 1937) as would be expected during compressive folding (Plate VII.6, 7).

With the exception of a few kink folds (B_2) developed in the foliation (S_1) west of Mt. Corbató, the entire area was folded and faulted during a single episode of deformation. No superimposed folding was anywhere observed.

Following deformation portions of the Duncan Mountains and Mt. Fairweather were intruded by post-tectonic granitic plutons.

Metamorphism appears to have been developed in two stages. The first is represented by biotite and muscovite which grew in the cleavage plane (S_1). The second stage presumably accompanied the main batholithic intrusion in the area and is represented by the development of a hornfels paragenesis. McGregor (1965) has shown that the mineral associations related to the thermal stage of metamorphism are characteristic of the hornblende-hornfels facies. Several occurrences are of note. Andalusite crystals to 20 cm in length, sometimes associated with cordierite, have grown in the schists near the fault northwest of Mt. Corbató. Fibrous sillimanite occurs in the Duncan Formation at places adjacent to intrusive stocks. Also, biotite porphyroblasts spot many portions of the Duncan Formation, as well as the more pelitic beds of the Fairweather Formation exposed on Spur #20.

The structural relationship, with Duncan Formation up-faulted against Fairweather Formation-Henson Marble, is repeated at Mt. Henson, but there the dip of the fault is opposite to that in the Duncan Mountains and the sense of movement is normal. The fault itself is concave toward the southwest giving a vertical trace on the south face of Mt. Henson and curving markedly toward the northwest along the north face (Plate VII.7, 8).

As in the Duncan Mountains the highest portion of the section is exposed adjacent to the fault. Here, at its type locality, the Henson Marble is deformed into a large drag fold near the ridge crest. Bedding is slightly truncated toward the foot of the ridge but both the Duncan Formation and Henson Marble are closely conformable across the fault. An additional outcrop of the Henson Marble occurs just to the south at The Tusk where it is asymmetrically folded in the same sense as on Mt. Henson.

At Mt. Henson the Fairweather Formation conformably underlies the Henson Marble to the west where it eventually gives way to the Speed Granite. Here the Fairweather Formation has been converted to hornblende, biotite, and calc-schists, and none of the sedimentary characteristics are preserved.

Nilsen Plateau

On the west flank of Nilsen Plateau the Wyatt Formation appears to intrude the LaGorce Formation which has been folded into a steep-limbed syncline with slight development of axial plane cleavage. Shear zones cut through portions of the Wyatt Formation which is otherwise quite massive and without internal structure. Metamorphism has resulted in the growth of biotite, chlorite and muscovite in the groundmass of the Wyatt Formation. Plagioclase phenocrysts are usually badly sericitized and near Lindstrøm Peak they are saussuritized. Metamorphic grade appears to be within the albite-epidote hornfels facies, but on the northern side of Moraine Canyon towards the main batholith amphibolitic gneisses are developed.

ROSS OROGEN: TECTONIC MODEL (BYRD GLACIER TO PENSACOLA MOUNTAINS)

The conclusions reached in the preceding chapters on the late Precambrian-Cambrian volcanic terrain exposed in the Queen Maud Mountains fill a large void in understanding of the Ross Orogen, and thus permit the formulation of a new model for its development which incorporates these findings with the conclusions of researchers in other portions of the Transantarctic Mountains. The proposed model is limited to the region between Byrd Glacier and the Pensacola Mountains, since some type of discontinuity beneath Byrd Glacier disrupts the pattern perceived along the range to the south and east. The usual north-south fold trend of the Byrd Group arcs sharply to the east just south of Byrd Glacier. Such a geometry could represent right lateral transcurrent movement in the basement beneath Byrd Glacier or alternatively compression of the Byrd Group against resistant crust to both the north and west. Rocks north of Byrd Glacier are foliated and non-foliated granites assumed to be related to the Cambro-Ordovician Granite Harbour Intrusives in the McMurdo Sound region (Haskell and others, 1965), but no radiometric dating has been done on these rocks. If these granitic rocks are thought of as having been offset toward the craton from the main batholithic axis exposed between Shackleton Glacier and the Horlick Mountains, such a left-lateral direction of movement is inconsistent with the sense of drag in the folds of the Byrd Group. An alternative might be that some of the rocks north of Byrd Glacier are portions of an older crystalline basement related to the Nimrod Group in the Miller and Geologists Ranges which produced a buttressing effect during folding of the Byrd Group.

An assumption basic to the paleogeography of this model, but not to the sequence of events, is the approximate throughgoing, linear distribution of rock groups from one end of the area to the other. The arcing boundary fault along the front of the range between the Duncan Mountains and Byrd Glacier (Robinson, 1964; Barrett, 1965) appears to cut obliquely across a set of linear, petrotectonic elements which in part disappear to the south beneath the overlying Beacon Supergroup and the polar ice plateau. The axis of magmatic activity, as indicated by volcanic extrusions and a continuous granitic batholith, is exposed from around Shackleton Glacier to the Thiel Mountains; whereas the sedimentary elements exposed between Ramsey and Byrd Glaciers more adjacent to the craton than the magmatic arc appear to re-emerge in the Pensacola Mountains.

On the weight of an overwhelming amount of evidence in the last decade demonstrating the association of magmatic and compressive mobile belts with zones of crustal subduction at plate margins, it seems

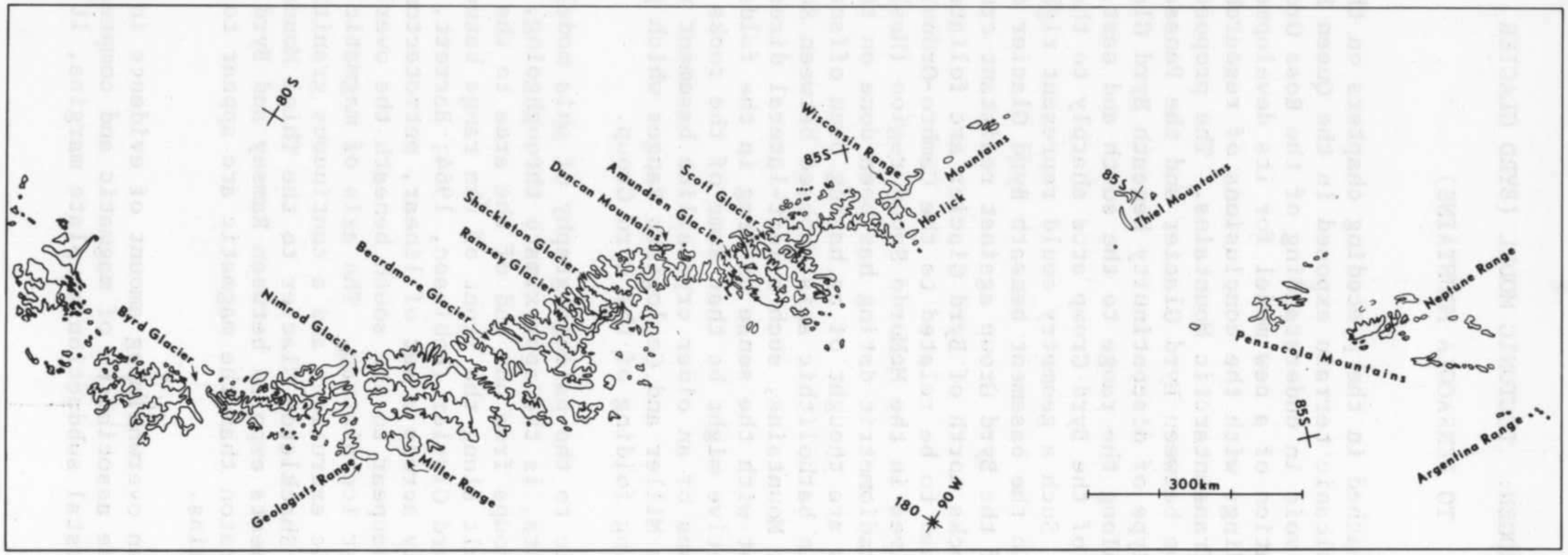


Fig. 9. Location map. Transantarctic Mountains - Byrd Glacier to the Pensacola Mountains

clear by analogy that subduction was occurring beneath the Pacific margin of Antarctica during the compressive phases of development of the Ross Orogen. Unfortunately the area where one would expect to find direct evidence of a subduction zone such as ophiolites, tectonic melanges, or blue schist is covered by ice from the Ross to the Filchner Ice Shelf.

Passive Phase

Following consolidation of the crystalline craton, Rb-Sr whole-rock isochron dated at 1.98 ± b.y. on the Nimrod Group (Gunner and Faure, 1972, graywacke-shale deposition commenced on the continental margin (Gunn and Warren, 1963; Grindley, 1963; Laird, 1964; McGregor, 1965; Murtaugh, 1969; Schmidt and others, 1964). In the Pensacola Mountains deposition began before emplacement of felsites (Schmidt and others, 1965), Rb-Sr whole-rock isochron dated at 1210 ± 76 m.y. (Easton, 1970). Rhyolites were emplaced at 1001 ± m.y. in the Bertrab and Littlewood Nunataks area 500 km east-northeast of the Pensacola Mountain on a portion of the craton which remained undeformed during succeeding events in the adjacent mobile belt. With the exception of the aforementioned volcanism, the continental margin appears to have been tectonically inactive in the late Precambrian throughout the period of flysch sedimentation.

Lying stratigraphically below the Goldie Formation and in a position close to the cratonic margin, alternating, well-bedded quartzitic and calcareous metasediments of the Cobham Formation may represent moderately shallow-water deposits in the Nimrod Glacier area (Laird and others, 1971) prior to further foundering of the continental terrace.

It is not known whether deposition commenced everywhere at once or whether it was active synchronously throughout the area at any time, but the entire margin did receive turbidite depositions, probably in an interconnected system of deep-sea fans, during the initial development of the Ross Orogen. (Formations include: Goldie Formation; Shackleton Coast area; Duncan Formation; Duncan Mountains area; LaGorce Formation; Amundsen-Reedy Glaciers area; Patuxent Formation; Pensacola Mountains).

Initial Compressive Phase (Beardmore Orogeny)

The passive regime became activated during the late Precambrian Beardmore Orogeny with the establishment of a calc-alkaline magmatic arc marginal to the continent and a zone of deformation and erosion between the arc and the craton. Porphyritic rhyolites and rhyodacites accumulated in voluminous deposits between the Thiel Mountains and Nilsen Plateau (Ford, 1964; Murtaugh, 1969; Minshew, 1967; this report). Based only on the single contact in the LaGorce Mountains the volcanic

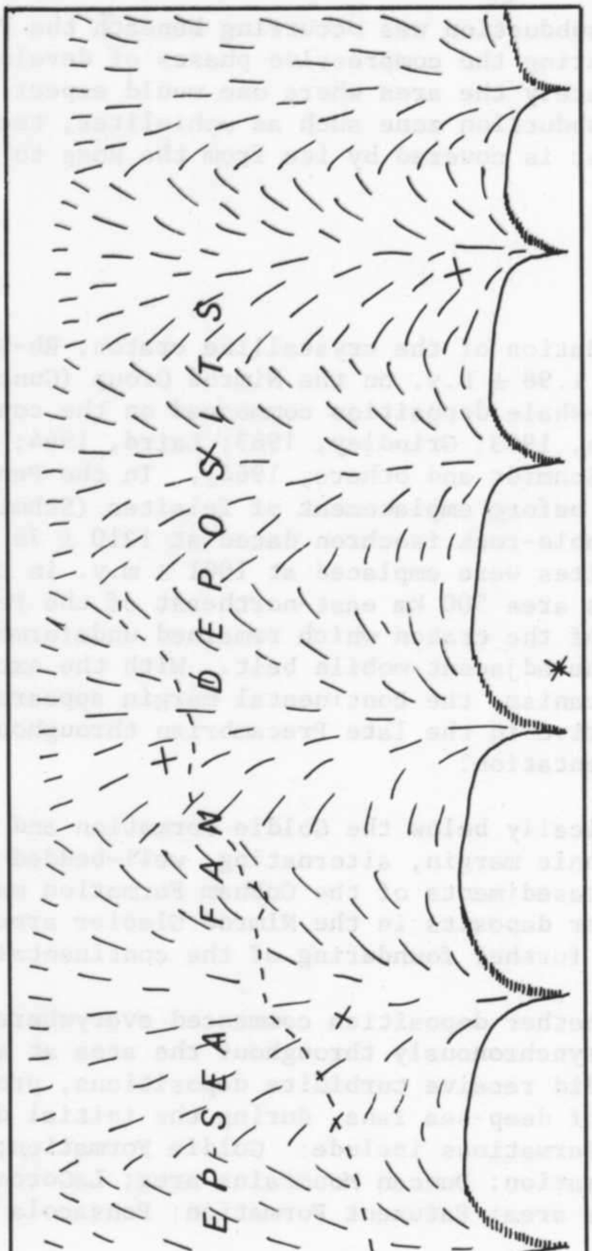


Figure 10. Ross Orogen - Passive Phase (Dashed line indicated coastline of Transantarctic Mountains)

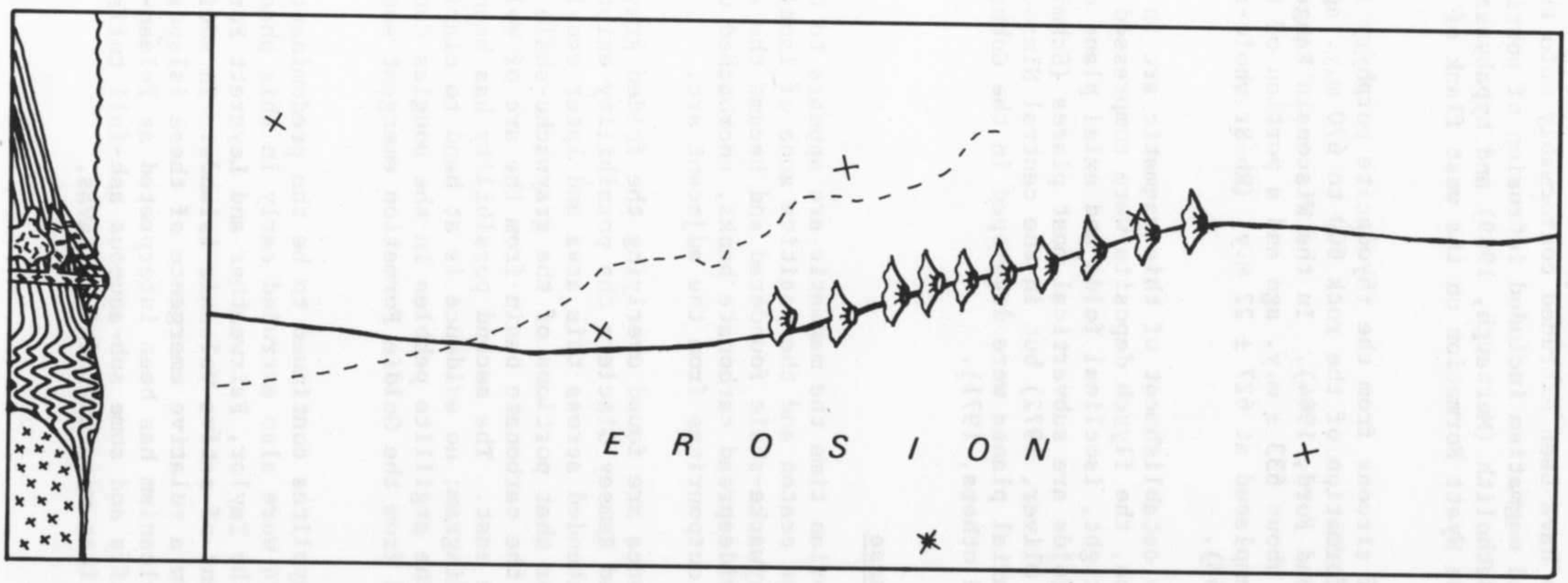


Figure 11. Ross Orogen - Initial Compressive Phase (Beardmore Orogeny) (Dashed line indicates coast-line of Transantarctic Mountains)

rocks are presumed to have been extruded conformably onto the underlying flysch. Related magmatism included intrusion of portions of the Wisconsin Range Batholith (Murtaugh, 1969) and hypabyssal emplacement of a phase of the Wyatt Formation on the west flank of Nilsen Plateau.

Lead-α dating of zircons from the rhyodacite porphyry in the Thiel Mountains indicated formation of the rock 600 to 670 m.y. ago (Ford and others, 1963; Aaron and Ford, 1964). In the Wisconsin Range the Wyatt Formation originated about 633 $\pm$ m.y. ago and a portion of the Wisconsin Range Batholith was emplaced at 627 $\pm$ 22 m.y. (Rb-Sr whole-rock isochrons, Faure and others, 1968).

Accompanying the establishment of this magmatic arc in a zone more proximal to the craton, the flysch deposits were compressed and uplifted with development of tight, isoclinal folds and axial plane cleavage. Axial planes of the folds are subvertical most places (Schmidt and others, 1965; Gunner, 1971a; Oliver, 1972) but in the central Nimrod Glacier area, subhorizontal axial planes were developed in the Cobham and Goldie Formations (Laird and others, 1971).

Middle Compressive Phase

By earliest Cambrian time the magmatic arc appears to have shifted slightly away from the craton and the positive zone of isoclinally folded and eroded graywacke-shale foundered and became the site of a back-arc basin with widespread carbonate banks, encroached upon occasionally by volcanic outpourings from the adjacent arc.

Since no limestones are found overlying the folded graywacke-shale between the Nimrod and Ramsey Glaciers the possibility exists that the back-arc basin extended across this area and later erosion removed any deposits there, or that portions of the graywacke-shale remained emergent, separating the carbonate basin from the arc of volcanic islands active to the east. The second possibility has been chosen arbitrarily for the diagram; no evidence is at hand to clarify the situation, although the argillite pebbles in the Douglas Conglomerate were probably derived from the Goldie Formation emergent somewhere in the region.

Calc-alkaline rhyolites continued to be the predominant volcanic rock type, but basalts were also extruded early in this phase of activity. Rocks of the Taylor, Fairweather and Leverett Formations indicate an environment of active volcanic islands. In addition local sequences seem to show a relative emergence of these islands through time. Much of the volcanism has been interpreted as Peléan-type producing ash-flow tuffs and some sub-aqueous ash-fall tuffs. Basalts and some of the rhyolites were extruded as lavas.

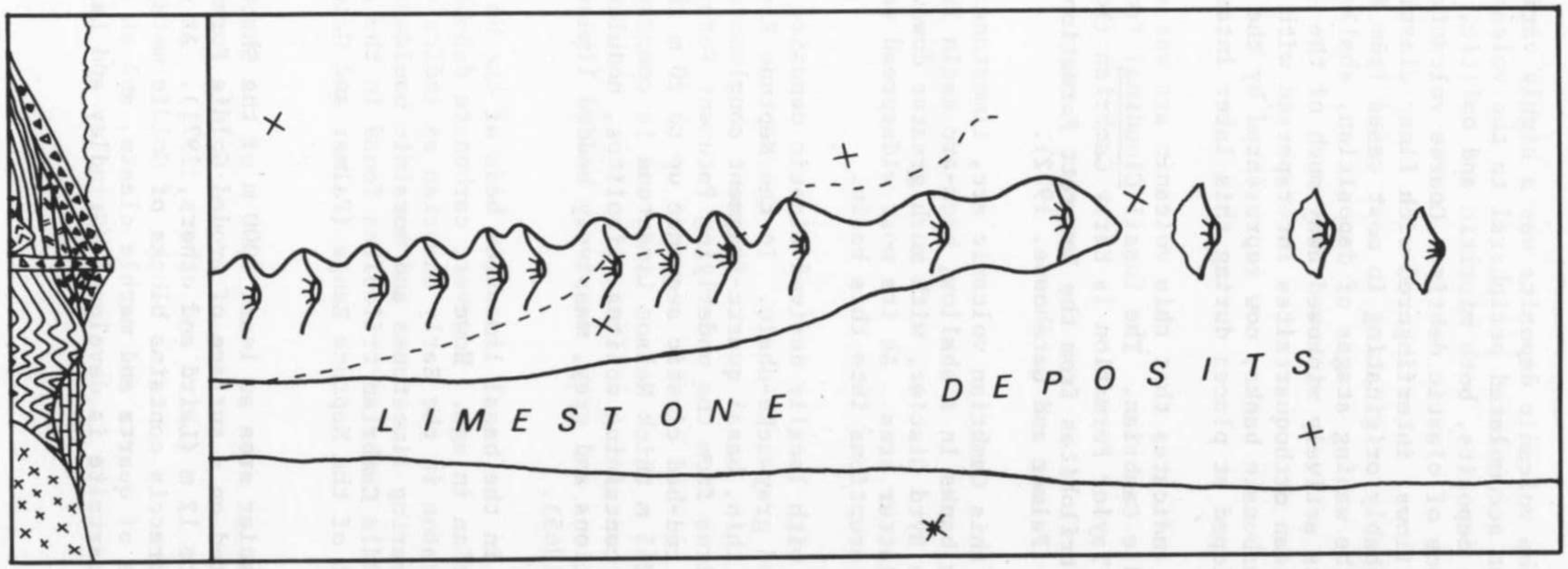


Figure 12. Ross Orogen - Middle Compressive Phase (Dashed line indicates coastline of Transantarctic Mountains)

Associated with the volcanic deposits was a highly varied assemblage of sediments that accumulated peripheral to the volcanoes. Included are carbonate deposits, both micritic and oolitic, and others with varying proportions of clastic debris. Coarse volcaniclastic influxes, due to mass flows, interfingered with finer clastic accumulations. "Chert," probably originating in most cases from fine ash, also formed. During the waning stages of deposition, shallow marine or terrestrial currents actively winnowed away much of the fine clastic fraction to produce clean orthoquartzites interspersed with intermittent volcanic deposits. Carbonate banks, now represented by the Henson Marble were also developed at places during this later interval.

Limited age data indicates that this volcanic arc was active during the Early and Middle Cambrian. The fossil Cloudina? from the middle portion of the Taylor Formation is Early Cambrian (Yochelson and Stump, (1977) and trilobites from the Leverett Formation appear to be Middle Cambrian (Palmer and Gatehouse, 1972).

Synchronous with this Cambrian volcanic arc, limestones were deposited on extensive banks in a shallow, back-arc basin from the Pensacola Mountains to Byrd Glacier, with much greater downwarp and sedimentation in the latter area. At its most widespread development the volcanic arc sent eruptions into this basin.

Deposition began with locally derived clastic deposits accumulating above the eroded graywacke-shale. In the Neptune Range of the Pensacola Mountains a thin, basal quartz-fragment conglomerate composed largely of detritus from the underlying Patuxent Formation, is overlain locally by a red-bed clastic sequence up to 20 m thick. The remainder of the 200-265 m thick Nelson Limestone is composed of gray thin-bedded limestone containing oolites, pisolites, nodules, and thin limey shale intercalations and gray, massively bedded limestones (Schmidt and others, 1965).

Trilobites found in the basal limestone beds of the Nelson Limestone are Middle Cambrian in age. However, carbonate deposition had begun closer to the craton in the Early Cambrian as indicated by in situ archaeocyathid-bearing limestones and morainic boulders containing Early Cambrian and Middle Cambrian trilobites found in the Argentina Range 250 km northeast of the Neptune Range (Palmer and Gatehouse, 1972).

In the Nimrod Glacier area at least 9000 m of the Shackleton Limestone were deposited on a surface of eroded Goldie Formation with relief undulating up to 12 m (Laird and others, 1971). At one locality a 1200 m thick basal breccia contains blocks of Goldie metagraywacke and a small percentage of quartz and marble clasts, and at another a 0-500 m thick basal quartzite is developed (Grindley and Laird, 1969).

Elsewhere limestone, sometimes containing sandy lenses, overlies the unconformity. The majority of the Shackleton Limestone is thick-bedded, gray to cream limestone, in part oolitic, with occasional interbedded lenses of quartz and limestone-pebble conglomerate and finer clastic rocks. Disharmonic, intraformational folds and coarse intraclasts indicate that slumping occurred locally. Fossils of Archaeocyatha from beds low in the section demonstrate that the formation had begun accumulating in the Early Cambrian (Laird and Waterhouse, 1962; Hill, 1964, 1965).

Laird and others (1971) have interpreted the environment of deposition for the Shackleton Limestone as a down-warping, shallow-water basin separated from an emergent coast to the west by an extensive archaeocyathid biocherm. Sedimentation in the basin usually kept pace with subsidence, but occasionally outpaced it, causing slumps.

Conglomeratic and less coarse clastic rocks occur on the northern and eastern margins of the limestone outcrops. Laird (1964) argued from the lithology of conglomeratic clasts that the Starshot Formation and Shackleton Limestone were approximately contemporaneous, and suggested correlation of the Starshot Formation with the Dick Formation and Douglas Conglomerate (Skinner, 1964). The Starshot Formation contains conglomerates with well-rounded cobbles of quartz and limestone, sandstones with ripple marks, current bedding and load casting, and shales (Laird, 1963). The Douglas Conglomerate likewise contains pebbles of quartz and limestone, but minor amounts of argillite, siltstone, quartzite, arkosic sandstone, diorite, granite and gneiss pebbles also occur (Skinner, 1965). Conformably underlying it is the Dick Formation composed of argillite and sandstone with grit lenses. The provenance of these formations is the local Shackleton Limestone, the underlying Goldie Formation and a plutonic-metamorphic terrain, probably the adjacent crystalline craton.

A spilite flow interbedded in the Dick Formation, and rhyolite and trachyte flows in the Starshot Formation likely represent the most distant volcanism from the magmatic arc to the east.

In the Pensacola Mountains area volcanic eruptions occurred locally over the Nelson Limestone. Rhyolitic flows, volcanic breccia deposits, hypabyssal intrusions, and clastic sediments of the Gambacorta Formation attain a thickness of 330 m and disappear completely within 3-5 km (Schmidt and others, 1965). The upper sediments intertongue with the shales, siltstones and fine sandstones of the overlying Weins Formation (Williams, 1969).

Culminating Compressive Phase (Ross Orogeny)

The main phase of the Ross Orogeny resulted in cessation of deposition, deformed all of the previously formed sedimentary rocks, and

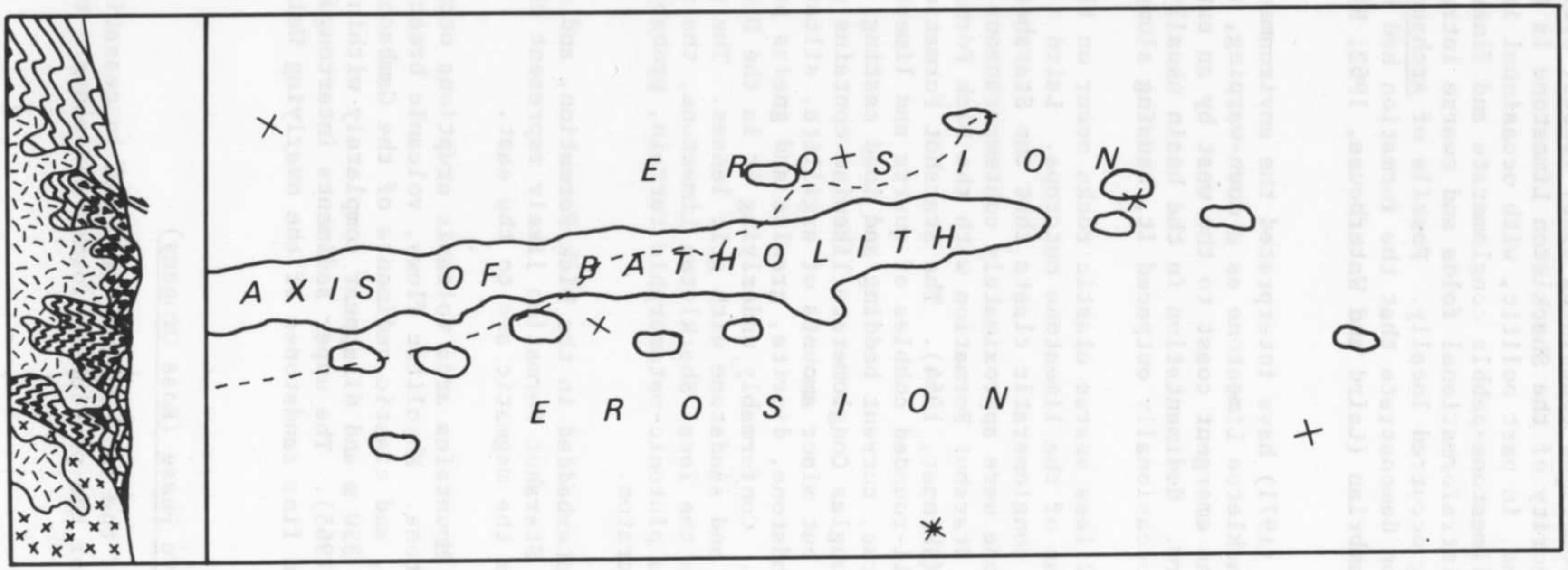


Figure 13. Ross Orogen - Culminating Compressive Phase (Ross Orogeny) (Dashed line indicates coast-line of Transantarctic Mountains)

produced widespread plutonism and metamorphism. Compression in the back-arc basin caused folding in the Cambrian limestones, more open and symmetrical in the Pensacola Mountains area (Schmidt and others, 1965) and tighter in the Nimrod-Byrd Glaciers area (Laird, 1963; Skinner, 1964), with local complexities occurring in both places. These folds evidentally developed through reactivation of structures in the underlying isoclinally folded graywacke-shale, for thus far no workers have reported intersecting fold trends in these rocks which would indicate a reorientation of stresses between the Beardmore and Ross Orogenies. Oliver (1972) does report crenulations and corrugations in addition to tight similar folds in the Goldie Formation near the mouth of Beardmore Glacier, but these appear to be local occurrences and not of the magnitude or abundance to argue for tectonic realignment.

Along the axis of the magmatic arc the downwarped sedimentary and volcanic rocks mobilized, producing complexly folded migmatites (Infrastructure).

Deformation, probably folding, of the higher, more rigid rocks (Superstructure) up-ended large sections of the former volcanic island complex, as exhibited in the Shackleton Glacier area. Mesoscopic folds and penetrative structures were developed only locally in these rocks. On the exterior side of the arc, regional metamorphism producing strong schistosity accompanied compression, which caused high-angle reverse faulting and asymmetric folding toward the craton.

Plutonism began before or during deformation in areas near the magmatic axis (McGregor, 1965; Burgener, in press; Murtaugh, 1969). But by far the most voluminous calc-alkaline intrusions arose after deformation, engulfing large portions of both the infra-structure and superstructure, and causing widespread hornfels-facies metamorphism (Gunn and Walcott, 1962; Ford and Aaron, 1962; Schmidt and others, 1965). The magmatic axis extends through the Queen Maud-Wisconsin Range Batholith from approximately Shackleton Glacier to the Horlick Mountains, disappearing beneath the Ross Ice Shelf and the polar ice cap. Anatectic conditions were reached in the migmatites intimately associated with the batholith at places around its margin. Toward the craton plutons become more isolated and sharply cross-cut the country rocks into which they were intruded. Gunner (1971a, 1974) concluded that the granitic rocks in the Beardmore Glacier area originated at least in part by melting of rocks similar in chemical and isotopic composition to the Nimrod and Beardmore Groups into which they intrude.

Numerous radiometric age dates on the intrusions and associated metamorphic rocks throughout the region spread between 520 and 450 m.y. indicating that plutonism had begun in the upper Cambrian and that the orogen had eroded and cooled sufficiently by the middle Ordovician for argon retention to have commenced (Aaron and Ford, 1964; Craddock and others, 1964; Eastin, 1970; Eastin and Faure, 1972; Faure and others, 1968; Grindley and McDougall, 1969; Gunner, 1971a; Gunner and Faure, 1972; Gunner and Mattinson, 1975; McDougall and Grindley, 1965; Minshew, 1965).

Erosion continued until the Devonian when the next cycle of sedimentation began once again to blanket the area.

Part II: Southern Africa

PRE-CAPE ROCKS, DESCRIPTIONS AND SUMMARY

Introduction

Rocks of pre-Cape age crop out in four windows through the Table Mountain Group in the Cape Mountains, South Africa. These include the Gamtoos area, the Cango area, the George area, and the Worcester-Swellendam area (Figure 14). The most extensive outcroppings of pre-Cape rocks occur to the west of Worcester, where they are called the Malmesbury Group (Figure 23, p.

Isotopic dating on the Cape Granite which intrudes the Malmesbury Group of the western Cape has yielded ages between 505 and 610 m.y. (Allsopp and Kolbe, 1965; Burger and Coertze, 1973). The concordant U-Pb date of 610 + 20 m.y. gives a lower limit for the age of the Malmesbury.

The pre-Cape rocks are unfossiliferous, highly deformed, lacking in distinctive lithologic horizons, deeply weathered, and economically unimportant. In the last 80 years they have been correlated with almost every major system of rocks in southern Africa, except the Cape and Karroo. A somewhat extended history of the "Malmesbury Problem" is presented in Appendix B.

Rocks from the four pre-Cape windows are systematically described in the succeeding sections, along with a short summary of the Malmesbury Group of the southwestern Cape Province following Hartnady and others (1974). Conclusions are drawn regarding depositional environments and interrelations between the various areas.

South African nomenclature has traditionally used "system," "series," stage," and "zone" when designating lithostratigraphic units. This practice is currently being abandoned to conform with international practice, but terminology has not been finalized for the pre-Cape rocks, so I will follow the usage of previous authors even though this introduces some minor inconsistencies.

Gamtoos Area

In the eastern Cape pre-Cape rocks crop out in a narrow belt north of the Gamtoos River. The entire sequence has been overturned toward the northeast with dips as low as 40°. Deformation during the Cape Orogeny produced considerable mesoscopic folding and cleavage in the pre-Cape rocks which accommodated themselves to the larger-scale folding of the more competent quartzites of the overlying Table Mountain Group. This area is unique by virtue of the apparently conformable relationship between the pre-Cape and Cape sedimentary rocks.

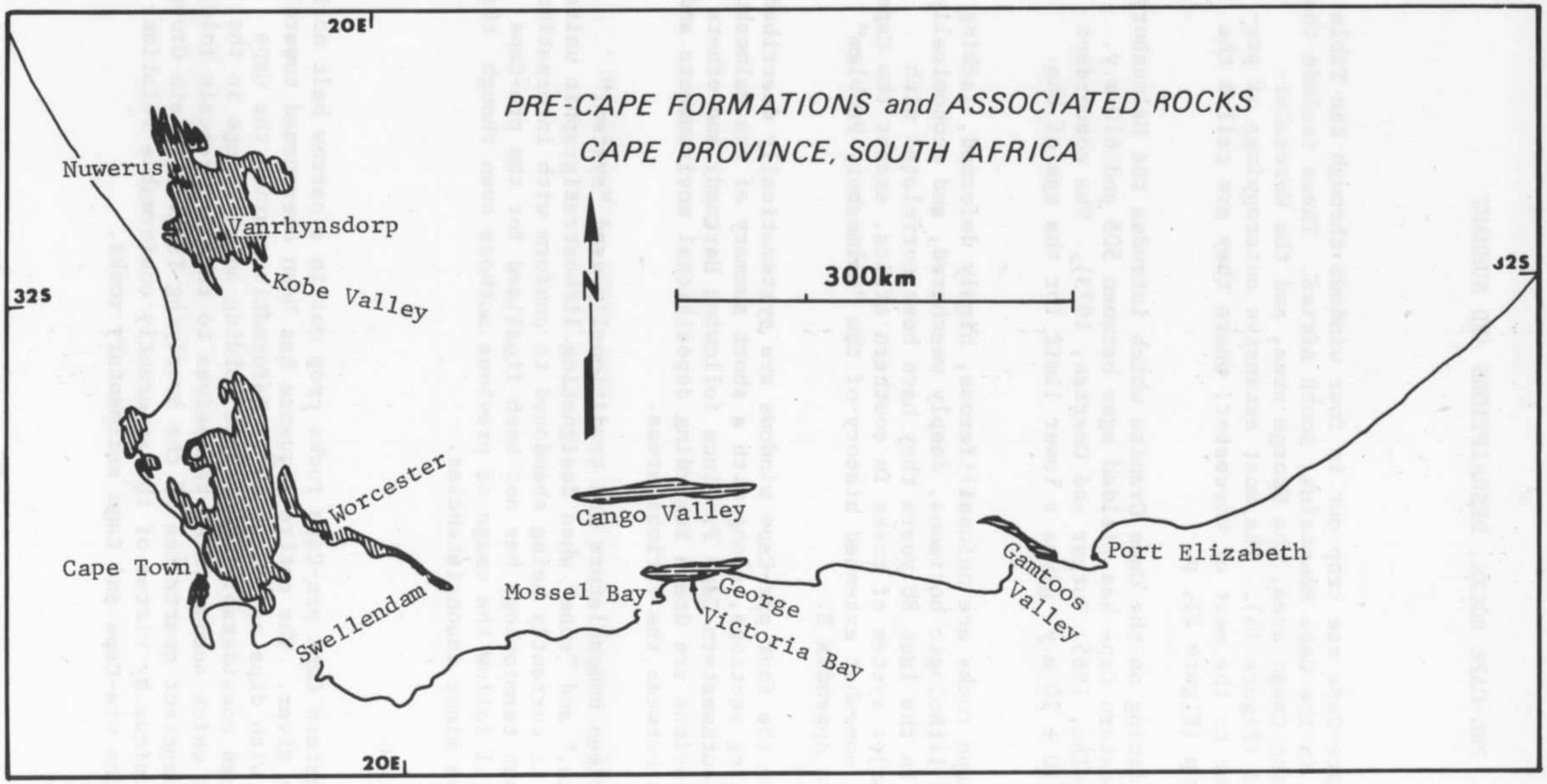


Figure 14. Pre-Cape formations and associated Rocks, Cape Province, South Africa.

The stratigraphic sequence was worked out by Amm (1934), Frankel (1936) and Haughton (1937). The final version, used on the Geological Survey's map sheet of the area is as follows:

Pre-Cape Stratigraphy-Gamtoos Area

Upper Pre-Cape	Arenaceous beds
	Kaan calcareous stage
Lower Pre-Cape	Phyllite and grit stage
	Kleinfontein calcareous stage
	Lower phyllite stage

Due to internal deformation and poor exposure the original authors did not estimate thicknesses, but judging from outcrop width on the map the pre-Cape strata may be as much as 4000 m thick.

The lower pre-Cape is characterized by two limestone horizons separated by phyllites and grits. The limestones are dark blue to gray. In thin section they can be recognized as sparites with scattered sub-angular to sub-rounded quartz grains. Thin beds of calcareous shale are also present at certain locations.

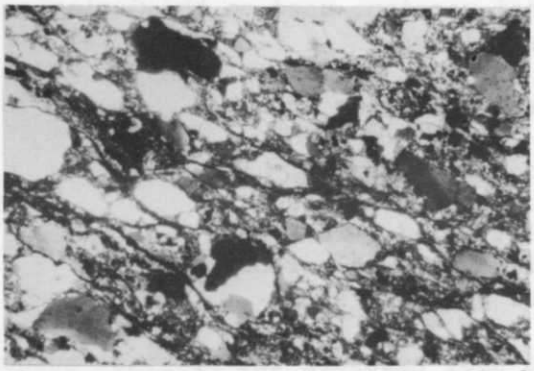
The Phyllite and grit stage is predominantly thin-bedded, greenish-gray, gray and tan phyllite, often highly cleaved. Interspersed through this are layers and lenses of micaceous quartzite, feldspathic grit and conglomerate (Plate VIII.1). The conglomerates are composed of pebbles and cobbles of vein quartz and quartzite, and occasionally clay galls are present.

Following the Kaan calcareous stage, the upper pre-Cape begins with a discontinuous basal conglomerate and passes up through a mixed sequence of micaceous quartzite, phyllite and grit becoming finer upwards and gradually losing the pelitic fraction with the development of ortho-quartzites that merge with the Table Mountain Group. Cross-bedding in the quartzites appears to increase in frequency higher in the section. Once again the grits and conglomerates are discontinuous lenses.

The quartzites are generally medium to very coarse-grained and usually contain a considerable fraction of feldspar, so that they may rightly be called subarkosic or occasionally arkosic (Plate VIII.2). K-felspar is more common than plagioclase, with microcline predominating over orthoclase. Detrial muscovite is also present in small amounts in some samples.

The quartz is ordinarily polygonized and highly sutured so that characteristics of the original grains are lost. The matrix is fine muscovite in the cleaner quartzites, but biotite or chlorite is developed as well in the more pelitic varieties. Magnetite, hemitite, zircon and tourmaline define laminations in the cross-bedding. The mineralogy of the clastic sediments indicates a crystalline (plutonic or metamorphic) provenance.

PLATE VIII – PRE-CAPE ROCKS



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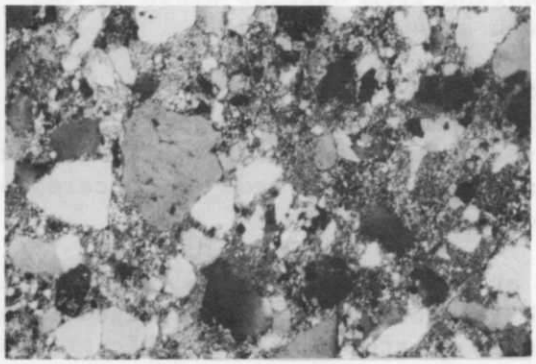

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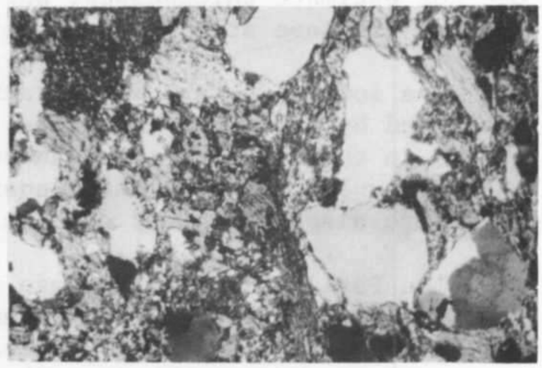

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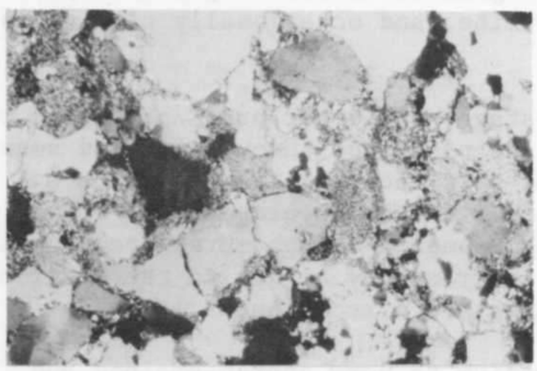

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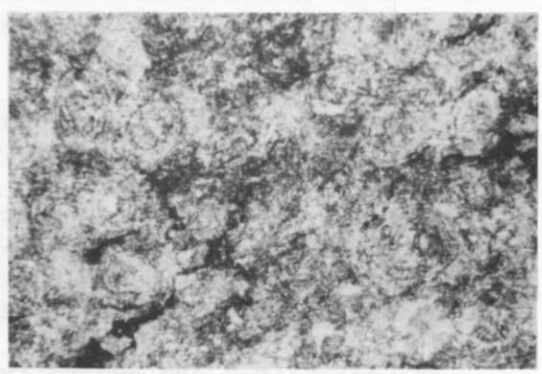

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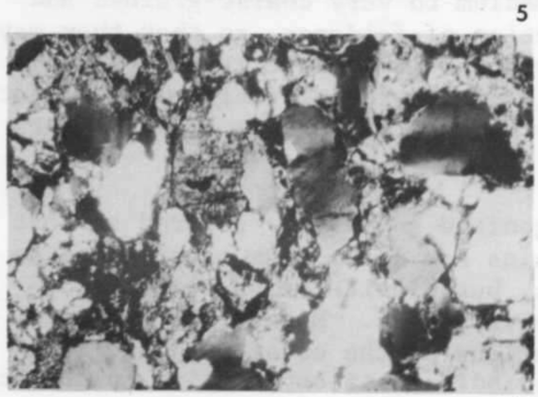

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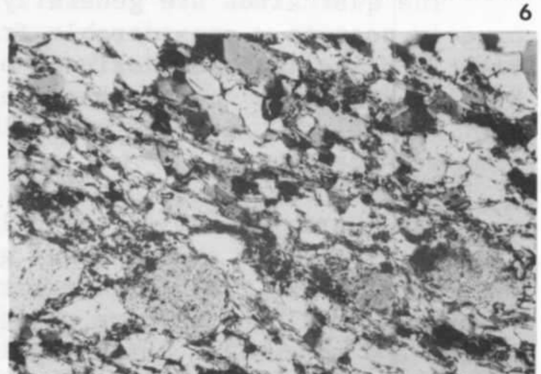

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Plate VIII - Pre-Cape Rocks

Plate VIII.1 Phyllite and Grit Stage, Gamtoos area. Quartzite with clasts of quartz, plagioclase and K-feldspar and matrix of biotite, muscovite and quartz. Crossed nicols, 25x. (AFH)

Plate VIII.2 Upper Pre-Cape, arenaceous beds, Gamtoos area. Quartzite with quartz, microcline and orthoclase. Crossed nicols, 63x. (AGD)

Plate VIII.3 Cross-bedded Grit Zone, Cango area. Quartzite with clasts of quartz and phyllitic rock fragments and matrix of chlorite and muscovite. Crossed nicols, 25x. (AEL)

Plate VII.4 Cross-bedded Grit Zone, Cango area. Clacareous quartzite with clasts of quartz, plagioclase, muscovite and quartzitic rock fragments and matrix of sparry calcite, chlorite and muscovite. Crossed nicols, 25x. (AET)

Plate VIII.5 Cross-bedded Grit Zone, Cango area. Quartzite with clasts of quartz and rock fragments and matrix of muscovite. Crossed nicols, 25x. (AFC)

Plate VIII.6 Limestone Zone, Cango area. Oosparite. Plane light, 25x. (AFE')

Plate VIII.7 Grit lens in Limestone Zone, Cango area. Quartzite with clasts of quartz, plagioclase and orthoclase and matrix of muscovite. Crossed nicols, 25x. (AEP)

Plate VIII.8 Victoria Bay phyllite, George area. Schist with plagioclase, quartz and muscovite. Crossed nicols, 63x. (ACH)

Interpretation

A brief interpretation of the depositional environment in this and each of the following areas will be attempted. These interpretations are admittedly based on limited data, and should be regarded with all proper reservation.

In the Gamtoos area during Lower pre-Cape time limestones accumulated in a shallow-marine shelf environment. This was interrupted for a time by the influx of deltaic deposits of the Phyllite and grit stage, the coarse-grained lenses representing channel and distributary mouth deposits.

Tectonic activity in the source area caused the invasion of coarse clastic sediments, again deposited in a deltaic system, (Upper pre-Cape). The source was distant enough or uplift limited such that a basal conglomerate was not everywhere developed. Through time stream competency was reduced and an active marine environment returned, winnowing away the silt and clay fraction and producing the mature arenites of the Table Mountain Group.

Cango Area

Pre-Cape rocks cropping out in an elongate belt in the Cango Valley are asymmetrically folded and overturned toward the north where they are overlain with angular unconformity by the higher Table Mountain Group. Cleavage is developed in portions of the Cango Group and some of the rocks are highly sheared, but in some areas the rocks retain very well their original bedding and sedimentary structures.

The principal contributions to the stratigraphy of these rocks have been by McIntyre (1932), Stocken (1954), and Mulder (1954). The nomenclature adopted by the Geological Survey (Roussouw and others, 1964) on their map sheet of the area closely followed that of Stocken (1954) and is used here.

Pre-Cape Stratigraphy-Cango Area

Upper Greywacke Zone
Cross-bedded Grit Zone
Lower Greywacke Zone
Limestone Zone

Again, thicknesses are difficult to determine due to structural complications, but the main limestone band in the Limestone Zone is in excess of 1300 m and the Cross-bedded Grit Zone appears to be about 2000 m thick (Roussouw and others, 1964).

The Cango Group is composed of limestone and clastic rocks including shale, graywacke, arenite, grit and conglomerate. In its lower portion the Limestone Zone contains intercalations of shale and limestone with subsidiary lenses of arkosic grit, graywacke, and arenite. This passes upward into the main limestone unit within which many solution caverns, including the Cango Cave, are developed.

The limestones are dark blue micrites and sparites. Some are massive, while others exhibit distinct lamination or layering, sometimes alternating with calcareous shale beds (10-50 cm thick). They are nearly pure carbonate, often with interspersed quartz grains (to 3 mm). Oolites with well developed concentric structure occur at occasional horizons, as do interclasts elsewhere (Plate VIII.6).

The shales are shades of gray, green and tan. Where unsheared they are often laminated. Gradations occur from the shale into coarser-grained siltstones, graywackes and arenites or into clacareous shale and limestone.

The lenses of grit grade into finer-grained clastic rocks or change abruptly at contacts. Sorting varies among the lenses but generally is rather poor. The angular, loosely-packed clasts include white and gray quartz, lesser amounts of microcline and orthoclase and minor plagioclase and perthite, surrounded by a pelitic matrix composed primarily of recrystallized muscovite and quartz.

The Lower Graywacke Zone contains graywacke and shale, as well as minor lenses of grit and limestone. Angular to sub-angular quartz and minor feldspar occur in a matrix of chlorite and quartz which makes up about 40% of the rock. Occurrences of this zone are severely deformed throughout most of the area.

Conformably overlying these rocks is the Cross-bedded Grit Zone. It begins with a basal conglomerate, first described by Corstorphine (1896a) and named the "Cango Conglomerate" by Rogers and Schwarz (1898b), and changes upwards to a sequence of cross-bedded feldspathic grits and arenites. Both monomict and polymict conglomerates are represented. The monomict variety contains well-rounded pebbles and cobbles of white quartz in a quartzitic matrix. Clasts of the polymict variety are primarily graywacke, arenite and shale similar in composition to the underlying sediments. Some quartz pebbles are present, as are occasional cobbles and boulders of granite and geniss. It was these crystalline clasts which led Corstorphine (1898) originally to suggest that the Cango Conglomerate was younger than the Malmesbury and cross-cutting Cape Granite of the western Cape. Unlike the clean arenaceous matrix of the monomict conglomerates, the polymict conglomerates contain a gritty, pelitic groundmass which sometimes exceeds 50% of the rock and completely separates the cobbles and boulders from one another. In the central portion of the area this matrix is highly sheared and the clasts have been rotated into the plane of deformation.

The conglomerates grade upward gradually into cross-bedded grits with only occasional lenses of conglomerate containing quartz pebbles and clay galls. The bluish and greenish-gray grits contain about 85% angular quartz grains with minor amounts of microcline, orthoclase and lesser amounts of plagioclase. A pelitic matrix of quartz, muscovite and sometimes chlorite makes up 10-15% of the rock (Plates VIII, 3,4,5).

The cross-beds are long (foresets to 3 m), low-angled (about 15^{o}) and invariably show direction of transport from the west (Stocken, 1954). Ripple marks were also observed on a few surfaces.

The Cross-bedded Grit Stage grades up into the Upper Graywacke Zone containing alternating layers of graywacke and shale which range in thickness from 1 m at the base to 3 cm at the top.

Interpretation

The lower portion of the Limestone Zone indicates a shallow-marine environment in which limestones and shales were deposited and across which currents at times carried coarse clastic debris. Conditions stabilized and a widespread carbonate shelf was established producing the main limestone unit of the Limestone Zone. This was followed by the introduction of more clastic material (Lower Graywacke Zone). Then tectonic uplift to the west caused the sudden influx of great amounts of coarse alluvial debris. The lenticular nature of the deposits, and sedimentary structures indicate a fluvial environment. As the tectonic highlands were reduced clast size in the Cross-bedded Grit Zone decreased, eventually giving way to the rhythmic alternations of the Upper Graywacke Zone interpreted as due to "seasonal flooding of a mature stream of waning competency" (Roussouw and others, 1964).

George Area

In the George area pre-Cape rocks surround the syntectonic George Granite. The rocks have suffered low grade regional metamorphism, probably not beyond the biotite stage, but thermal effects related to the intrusion have produced andalusite and cordierite locally. The one systematic treatment of this area was by Potgieter (1950b), who determined the stratigraphy and described the petrography of the rocks. Occurrences are considerably deformed so that his thickness data, particularly in the Homtini phyllites and basal argillaceous horizon, are only approximate:

Pre-Cape Stratigraphy--George Area

	Approx. thickness in feet
Homtini phyllites	3,700
Victoria Bay feldspathic quartzites	2,200
Victoria Bay phyllites.	1,000
Kaaimansgat quartzites.	970
Kaaimansgat phyllite band	20
Groot Hoek quartz schist.	2,660
Basal argillaceous horizon . . . . . at least	3,000
Total thickness . . . . . . at least	13,550 ft.

The basal argillaceous horizon consists largely of gray and tan shale, siltstone, phyllite, and minor amounts of graywacke, recrystallized to schists. The rocks are usually sheared and contorted, but at places they appear to be thinly bedded and laminated. The graywacke contains angular quartz, appreciable twinned plagioclase, and occasional detrital muscovite. The grains are poorly sorted and are surrounded by a considerable amount of matrix material consisting of recrystallized quartz, muscovite and biotite. Due to this recrystallization it is impossible to distinguish the finer clasts from the recrystallized quartz of the matrix.

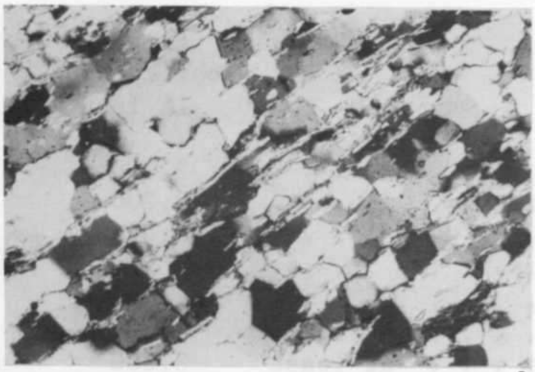
The Groot Hoek quartz schists are highly arenacous rocks containing a relatively large amount of muscovite and sericite. They are evenly bedded and in most cases have a sharp contact with the underlying argillites.

Conformably overlying the quartz schists is the Kaaimansgat phyllite band, a distinctive unit composed largely of muscovite and biotite surrounding small (0.1 mm) quartz grains and accessory feldspar.

This is followed by the Kaaimansgat quartzite, a medium-grained, fairly massive unit composed primarily of recrystallized quartz, with muscovite and a little biotite (Plate IX.1). Both microcline and plagioclase occur in some portions. An argillaceous horizon closely resembling the Kaaimansgat phyllite band, is discontinuously developed within this unit.

The Victoria Bay phyllites follow conformably. They are a sequence of light and dark colored phyllite and schist (Plate VIII.8). Muscovite is the predominant mica in many of the layers, but biotite exceeds it occasionally. More arenaceous beds, containing quartz and

PLATE IX — PRE-CAPE ROCKS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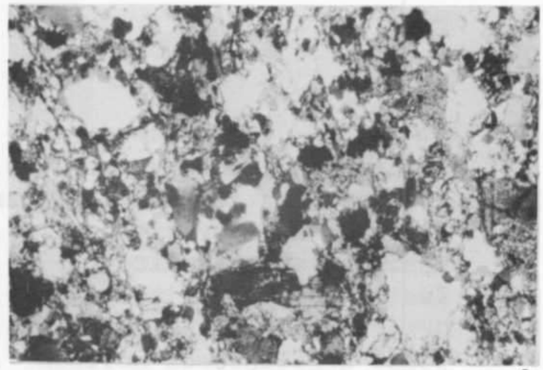
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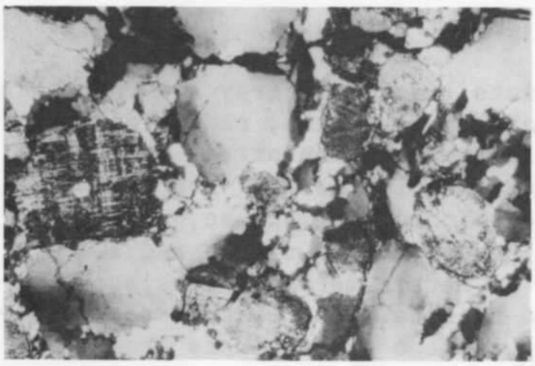
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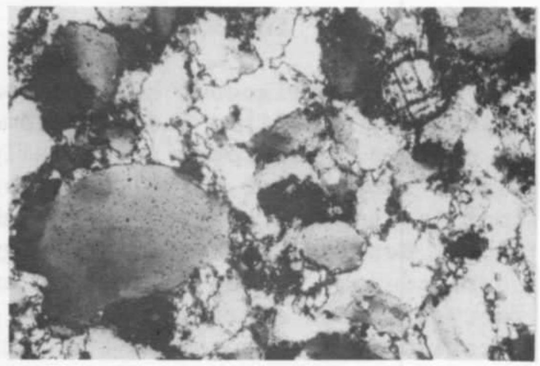
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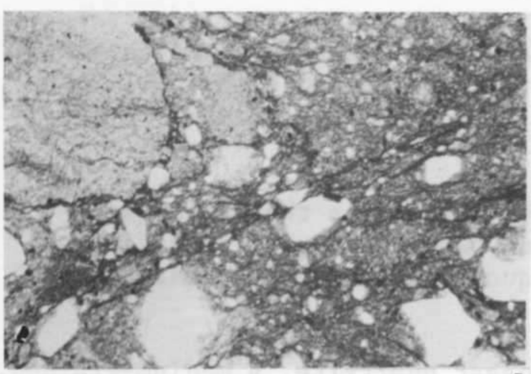
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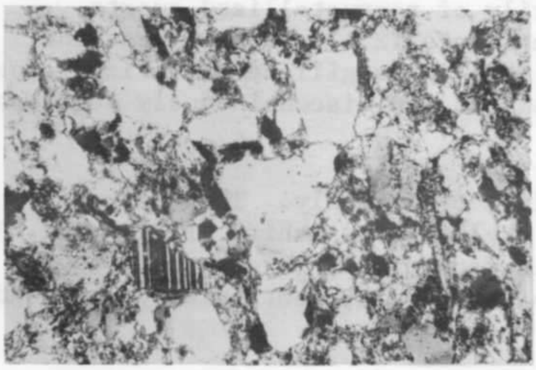
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Plate IX - Pre-Cape Rocks

Plate IX.1	Kaaimansgat quartzite, Goerge area. Quartz, muscovite and biotite. Crossed nicols, 63x. (ACI)
Plate IX.2	Coarse-grained lens in Homtini phyllite, George area. Quartzite with clasts of quartz, plagioclase and muscovite and matrix of muscovite, biotite and quartz. Crossed nicols, 25x. (ACO)
Plate IX.3	Victoria Bay feldspathic quartzite, George area. Quartzite with clasts of quartz, microcline, orthoclase and plagioclase and minor matrix of muscovite. Crossed nicols, 25x. (ACV)
Plate IX.4	Porterville Formation, Worcester area. Quartzite with clasts of quartz and microcline and minor matrix of biotite and chlorite. Crossed nicols, 63x. (ADE)
Plate IX.5	Porterville Formation, Worcester area. Breccia. Plane light, 25x, side view. (ADJ)
Plate IX.6	North of Robertson. Laminated marble with tremolite needles. Plane light, 25x. (ABR)
Plate IX.7	Tygerberg Formation, quarry northeast of Killarny Race Track, north of Cape Town. Graywacke with clasts of quartz, plagioclase and muscovite and matrix of sericite, biotite, quartz and plagioclase(?). Crossed nicols, 63x. (AAY)
Plate IX.8	Folded quartzite layer in Porterville Formation, northwest of Worcester.

plagioclase with minor microcline also occur at places, alternating at 5 to 30 cm intervals with the phyllitic material.

Overlying the phyllites are the Victoria Bay quartzites, a sequence of relatively clean, coarse-grained quartzites with gritty and conglomeratic lenses (Plate IX.3). Quartz is often recrystallized. Microcline, which appears to be sub-angular to sub-rounded, is the predominant feldspar, but twinned plagioclase is also present. Muscovite is developed only slightly in the quartzites, whereas in the gritty lenses a peltic matrix of muscovite, biotite and quartz is conspicuous. In the area of Victoria Bay numerous, thin (5-20) marble beds are developed in the quartzites.

The uppermost rocks in the George area are the Homtini phyllites, which consist of an alternating sequence of bluish and greenish-gray slates, phyllites and metagraywackes, alternating with lenses of quartzite and grit. Bedding varies from 20 cm to 1 m, the phyllites usually being on the thinner side and laminated. The coarser material consists of poorly sorted, sub-angular grains set in a matrix of quartz, muscovite and a little biotite (Plate IX.2). Quartz predominates but feldspar is important, with plagioclase exceeding microcline and orthoclase. Detrital muscovite is a conspicuous accessory mineral and phyllitic rock fragments are abundant in some places.

Interpretation

The metamorphic overprint in the George area makes interpretation difficult. It can be said, however, that the fine-grained, pelitic rocks of the lower section represent low-energy, probably deeper water conditions, and that the Homtini phyllites with their lenses of coarser clastic material signal emergence and the influx of channeled deposits, as would occur in a deltaic environment.

Wocester-Swellendam Area

The fourth group of pre-Cape rocks to be considered here crop out in a narrow strip from Swellendam to west of Worcester where they connect with outcrops of the Malmesbury Group in the vicinity of its type area, recently summarized by Hartnady and others (1974). Unlike the other pre-Cape windows farther to the east, no comprehensive studies have been undertaken here. The area was geologically mapped by workers at the University of Stellenbosch (1948); but, whereas the map is useful for showing the distribution of rock types, a consistent stratigraphy was not worked out nor was any written report produced. On the other hand various geologists have studied selected areas within the Worcester-Swellendam strip and described local sequences and structure (Schwarz, 1896, 1897a, 1905b; Rastall, 1911; de Jager, 1941; de Villiers and others, 1964; Hartnady, 1969).

The Brandwacht Formation crops out north of Worcester where it overlies the Porterville Formation extensively developed to the north and west in the area including Tulbagh, Hermon and Porterville (Hartnady and others, 1974). East of the Nuy River the pre-Cape rocks exhibit a variety of rock types which appear similar to those of the Porterville Formation although further study may determine that some of these are representative of the Brandwacht Formations.

The Porterville Formation in the Worcester-Swellendam area is composed of phyllite and fine to medium-grained graywacke with numerous intercalations of coarse-grained arenite and some limestone. The phyllites and graywackes are bluish or greenish gray when fresh, but most have been deeply weathered and are shades of tan or buff. Much of the area has been highly sheared so that original textures and bedding relations are destroyed. However, where intact the rocks are usually bedded at 5-20 cm intervals, although coarse-grained units may be as thick as 1 m. The phyllites are often laminated, but as has been pointed out by Hartnady and others (1974), there do not appear to be graded beds.

Between Robertson and Ashton are numerous gritty and conglomeritic lenses containing considerable amounts of pelitic matrix. The conglomerates are often stretched and may contain up to boulder-size clasts of vein quartz, quartzite, and graywacke. In addition, occasional pebbles and cobbles of granite were found in deposits near Klaas Voogd's Rivier, east-northeast of Robertson. The gritty arenaceous beds usually contain poorly sorted, sub-angular quartz, plagioclase, K-feldspar, and sometimes phyllitic rock fragments, although beds containing only quartz clasts were also found. The quartz is often a mixture of white and bluish-gray varieties similar to that described in gritty quartzites of the Porterville Formation between Voelvlei and Wellington (Hartnady and others, 1974). Both microcline and orthoclase are usually present in varying proportions. Calcareous arenites also occur at places east of Robertson with quartz grains set in a sparry matrix composing 30-40% of the rock.

Between Ashton and Swellendam conglomerates are not developed and the quartzite which is found in sporadic lenses tends to be cleaner than the more pelitic varieties abundantly developed toward Robertson.

Numerous limestone lenses are developed in the vicinity of Robertson and a few minor occurrences persist almost to Swellendam. The limestones are dark bluish-gray or light gray and may be either finely bedded or massive (Plate IX.6). Both micrite and sparite occur, but the latter is possibly due to recrystallization related to metamorphism or deformation. Isolated, sub-angular quartz grains are characteristic of some of these deposits, while others are quite pure carbonate.

In the vicinity of Hex River and Brandwacht, the Porterville Formation (Glen Heatlie Formation of Hartnady, 1969) occurs as gray-green phyllites and graywackes beneath the Brandwacht Formation.

Around Waaihoek the Porterville Formation (Waaihoek Formation of Hartnady, 1969) contains numerous beds of conglomerate and quartzite interbedded with the characteristic phyllite and graywacke. Clasts range up to boulder-size in the conglomerates and are composed of quartz, quartzite and some phyllite. Sorting is moderate in some occurrences but in others the rocks are a chaotic jumble of quartz and phyllitic rock fragments, having the appearance of a locally derived slump deposit (Plate, IX.5). Certain of the quartzites contain considerable pelitic matrix, but more often they are quite clean with moderately to well-sorted, sub-angular to sub-rounded quartz grains. Little or no feldspar occurs in these beds.

The Brandwacht Formation overlies the Porterville Formation north of Worcester, and perhaps is represented by some of the coarser units in the vicinity of Robertson. It is composed of greenish-gray and gray graywackes and phyllites with a basal conglomerate (to 100 m) and numerous conglomeratic lenses higher in the section. The conglomerates are a jumble of graywacke and phyllite chips and fragments plus lesser quartz pebbles. Compositions of the fragments are similar to the rocks in which they are intercalated and derivation appears to have been local.

Interpretation

The lack of stratigraphic control in this area hinders interpretation. Nevertheless, the assembled lithologies are similar in many ways to those in the Cango and Gamtoos areas. The clean quartzite and limestone lenses are characteristic of marine shelf conditions during at least part of the deposition of the Porterville Formation. However, the pelitic grits and conglomerates found around Robertson and Waaihoek once again indicate the introduction of coarse-clastic channel deposits.

Tectonic unrest within the immediate vicinity caused downwarping of a local basin which received graywackes and shales as well as the conglomeratic slump deposits of the Brandwacht Formation.

Southwestern Cape

Rocks in the southwestern Cape have recently been summarized by Hartnady and others (1974). They distinguish three tectonic domains, separated by narrow belts of major dislocation, each with a characteristic assemblage of formations. The northeastern tectonic domain which includes the area as far east as Worcester contains rocks of the Boland

Subgroup, similar to those described in the preceding section, as well as a more conglomeratic phase near Picketb e rg (see Figure 23). The central and southwestern tectonic domains (from Cape Town to Saldana to west of Piketberg to Wellington) contain primarily finely-bedded pelite and massively-bedded graywacke and impure quartzite of the Morreesburg and Tygerberg Formations (Plate IX.7).

Hartnady and others (1974) have suggested that the rocks of the southwestern and central tectonic domains were of turbidite origin and accumulated on a continental rise, and that rocks of the northeastern tectonic domain were of a shallower water environment, perhaps a continental shelf.

Summary

The closest similarities among the pre-Cape rocks exist between the Cango and Gamtoos areas where both lithologies and stratigraphy are closely comparable. Both areas contain prominent limestone horizons associated with phyllite, graywacke and arkosic grit in their lower portions, and both contain an upper cross-bedded clastic succession which begins with a basal conglomerate and becomes finer-grained up section. That the basal conglomerate and succeeding conglomeratic lenses are more extensively developed in the Cango may be explained by that area being closer than the Gamtoos to an uplifted source to the west, a situation compatible with cross-bed directions in the Cross-bedded Grit Zone of the Cango (Stocken, 1954).

The lack of an unconformity between the pre-Cape rocks and Table Mountain Group in the Gamtoos area also indicates that that area was beyond the limits of the tectonic zone which eventually encompassed all the pre-Cape rocks exposed elsewhere in the Cape.

As has been argued by Hartnady (1969), rocks of the Boland Subgroup around Worcester show broad similarities to those of the Cango. Lithologies of the Porterville Formation are comparable to the lower rocks in the Cango and Gamtoos areas, although limestones appear to be relatively less developed and coarse clastic rocks relatively more so.

The overlying Brandwacht Formation is one of locally derived slump deposits, whereas the Cross-bedded Grit Zone of the Cango and the Upper pre-Cape of the Gamtoos area appear to have been delivered by stream action, so that a strict lithologic comparison is not warranted. An interpretation compatible with what evidence exists is that the Brandwacht Formation accumulated in a very local basin at or near the uplifted source which supplied the conglomerates to the areas farther to the east.

The relationship of the Boland Subgroup to the Swartland Subgroup and Tygerberg Formation cropping out in tectonic domains to the west is still uncertain, but Hartnady and others (1974) suggest that these latter rocks may represent accumulations in a deeper-water facies perhaps analogous to a continental rise.

In attempting to compare the rocks of the Geroge area to those of the Cango or Worcester-Swellendam areas the only similar feature is that lenses of coarse material find their way into the upper portion of the succession (Homtini phyllites); but even so these are not nearly as well developed as the conglomerates of the other areas. Judging from the generally even bedding and fine-grained, pelitic nature of most of the rocks around George, they originated in a deep-water environment and are probably more comparable to rocks of the Swartland Subgroup and Tygerberg Formation of the western Cape.

If, as this analysis seems to indicate, there was a related influx of coarse clastic sedimentation in the upper portion of the sections in the Gamtoos, Cango and Worcester areas, this event may be related to the first upheaval of the Damara Orogeny as felt in the western Cape. There is no direct way of dating this event since the formations have thus far yielded no fossils, but the 610 ± 20 m.y. date of the Cape Granites (Burger and Coertze, 1973) indicates a time at which plutonism was in progress and suggests that the first deformation was latest Precambrian, perhaps slightly before the formation of the granite.

The main phase of the Damara Orogeny ceased sedimentation and caused deformation and erosion in all of the Cape except in the Gamtoos area. The Klipheuwel Formation represents local, rapid molasse sedimentation of the lithologically mature Table Mountain Group.

DAMARA OROGEN (SOUTH WEST AFRICA-SOUTH AFRICA):

SUMMARY AND CONCLUSIONS

Damara Supergroup

A belt of late Precambrian geosynclinal rocks occurs intermittently along the west coast of Africa from Gabon to the Cape, and includes the pre-Cape rocks discussed in the preceding chapter. In central and northern South West Africa they belong to the Damara Supergroup, (Kröner, 1974). The axis of sedimentation and orogenesis arcs sharply from both the north and south and strikes toward the interior of the continent in an east-northeast direction to form a two-sided mobile belt whose movement was directed outward from the central region (Figure 15). Each side of the basin contains both similar and distinctive features of stratigraphic and structural development. Geanticlinal ridges of older crystalline basement rocks have been active on both sides as loci dividing cratonic or shelf sediments from the deeper central portions of the basin. The northern part of the area exhibits all the elements of the classical geosyncline (Martin, 1965; Guj, 1970).

Present day terminology designates the northern miogeosynclinal rocks as the Otavi Group, and the southern and western eugeosynclinal rocks as the Swakop Group of the Damara Supergroup (Kroner, 1974). The Otavi Group is characterized by shallow-water carbonate rocks, a lack of volcanic rocks, slight if any metamorphism, and gentle folding. By contrast, the Swakop Group is a clastic sequence with some volcanic intercalations, which has been strongly deformed, locally intruded, and metamorphosed, in places to anatectic levels.

Along the southern fringe of the geosyncline the Swakop Group thins toward a basement high, beyond which are found the cratonic sedimentary rocks of the Nama System extending southward nearly to the Orange River (Germs, 1974). Folding and thrusting are intense north of this basement high (Guj, 1967) while immediately south of it the Nama occurs in large nappe sheets in the Nankluft Mountains (Korn and Martin, 1959; Martin, 1974) and in open folds more to the east.

The following outlines the sedimentary and tectonic development of the Damara Supergroup (see Table 7).

The arenaceous beds of the Nosib Group deposited upon a major unconformity record the first sedimentation in the Damara geosyncline. They are developed throughout the region, beneath both the Otavi and Swakop Groups, with the exception of the Welwitchia area (Frets, 1969) where the first deposition was that of the Otavi Group. Cross-bedded,

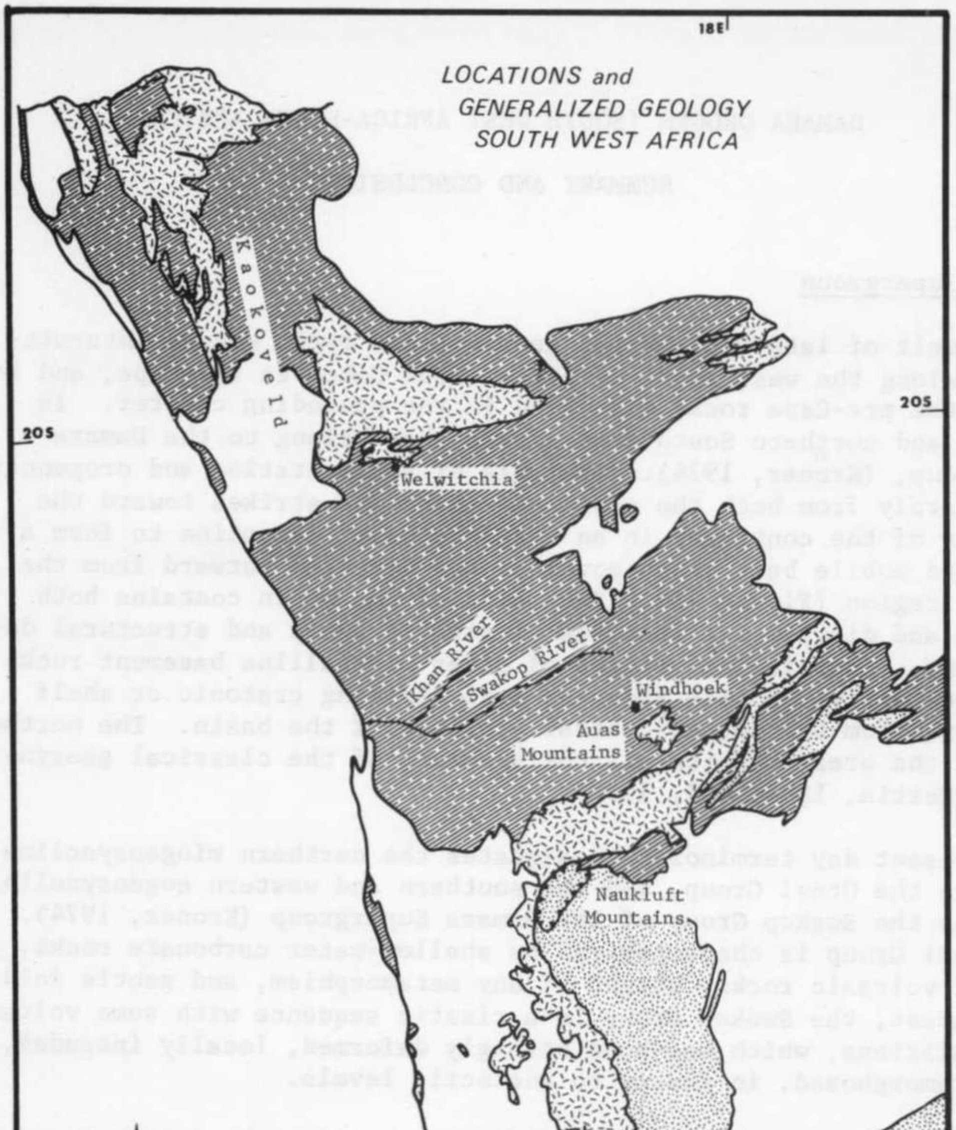


Figure 15. Locations and generalized geology, South West Africa (after Martin, 1965)

Table 7: Correlations of the Damara Supergroup

Kaokoveld Guj(1970)				Khan and Swakop Rivers Area Smith(1965)		Windhoek Area Schalk(1961) in Martin(1965)		Southern Margin Schalk (1970)
Mulden		Khomas		Khomas		Khomas		Nama
Otavi	L. Dolomite	Hakos	U. Hakos	Hakos	U. Hakos	Hakos	U. Hakos	
	Tillite		Tillite		Chuos Tillite		Tillite	Blaubeker (tilloid)
	U. Dolomite		L. Hakos		L. Hakos		L. Hakos	
Nosib		Nosib		Nosib		Nosib		Nosib

feldspathic quartzite, arkose and grit or their more metamorphic equivalents are the most common rock types, with minor occurrences of limestone, dolomite, and mafic lava recorded at places. Basal conglomerates are present at the southern margin of the sedimentary basin (Schalk, 1970) and locally elsewhere, while conglomeratic lenses are a minor constituent scattered throughout the sequence.

Deposition seems to have been in isolated basins with fluvial or shallow-marine environments, established on a pre-Damara landscape of considerable relief. As the regional trough developed these separate basins coalesced, with great and varying thicknesses of the Nosib Group accumulating throughout much of the region, although certain basement highs continued to remain as source areas for the duration of deposition (Martin, 1965).

An unconformity separates the Nosib Group from the overlying Otavi or Swakop Group in all but the Khan and Swakop Rivers area located in the central portion of the basin (Smith, 1965). However, its development is not complete in certain local areas (e.g., the southern Kaokoveld, Guj, 1970). Folds related to this period of activity are recorded on both the northern and southern margins of the orogen (Guj, 1970; Schalk, 1970). Metamorphism reaching middle amphibolite grade is recorded locally in the southern Kaokoveld (Guj, 1970) and some degree of metamorphism at this time is indicated in the area south of the Auas Mountains (Schalk, 1970, p. 39). This deformational and metamorphic event is sometimes correlated with the Katangan orogeny (580-660 m.y.) (Clifford, 1967).

Following this episode the miogeosynclinal and eugeosynclinal facies were established during renewed deposition. The carbonate rocks of the Otavi Group represent stable, platform sedimentation, while the Swakop Group including the calcareous Hakos Subgroup and overlying pelitic Khomas Subgroup is thought to represent increasingly unstable and subsiding trough conditions. A distinctive glaciogenic horizon in the Otavi and Khomas Groups serves as the one good marker of synchronous deposition in the sequence.

The lower portion of the Otavi Group is composed of well-stratified limestone and dolomite, not uncommonly containing stromatolites. Shales and some quartzite are also present. In the Lower Hakos Series one finds an association of marble, sometimes dolomitic, and pelitic schist with lesser amounts of quartzite, chert, and at places itabirite.

The tilloid horizon follows a marked discordance, in places an unconformity, throughout the area. It was named the Chuos Tillite by Gevers (1931) for rocks in the central region. There deposits of a basal moraine were postulated. Other occurrences point more towards a glaciomarine origin (Kroner and Rankama, 1972). This horizon also contains a peculiar association of hematite and hematite-magnetite deposits, well stratified but not banded (Martin, 1965).

The glacial event was succeeded by renewed carbonate deposition in both the Otavi and Swakop Groups. The upper portion of the Otavi Group is well-stratified micritic dolomite with widespread stromatolite horizons. In the southern Kaokoveld they reach reef proportions in the lower part of the sequence. The upper portion of the Hakos Subgroup is composed of calcareous and dolomitic marbles as well as interbeds of quartzite and pelitic schist. This is followed by the Khomas Subgroup, a thick, monotonous sequence of pelitic schist and schistose graywacke, with minor intercalations of quartzite and of marble at places (Gevers, 1934; Guj, 1970).

The final deposits of the northern area are known as the Mulden Group, a sequence of immature feldspathis quartzite, arkose, graywacke and phyllite, with conglomerate and calcarenite development in its lower portions toward the west. It rests with an angular unconformity on the Otavi Group in the western region of the basin, but towards the east the sequence shows a conformable relationship (Söhnge, 1964). Cross-beds and ripple marks as well as the lithologic association, indicate shallow marine or non-marine sedimentation. This formation has been interpreted as a molasse foredeep or exogeosynclinal deposit derived from a rising tectonic hinterland to the west during the initial activity of the Damara Orogeny (Frets, 1969; Guj, 1970).

The relationship of the Mulden and Khomas Groups has prompted recent debate. They are probably everwhere separated by the ridge system of basement highs. Martin (1965) arguing from limited field data felt that they should be correlated due to an apparent inter-fingering of the two groups (his Transition Facies) in the Welwitchia area. Frets (1969) later mapped this critical area in detail and concluded that all of the beds in question belonged to the Khomas Subgroup and were involved in the initial phase of folding that led to the deposition of the Mulden Group. However he stopped short of denying that all of the Khomas beds in all other areas also predated the Mulden Group and acknowledged that intraformational breaks corresponding to the unconformity beneath the Mulden Group could exist elsewhere in the Khomas; but, if so, are obscured due to subsequent tectonism.

Along the southern margin of the orogen the only post-Nosib rocks of the Damara System are the coarse clastic deposits of the Blaubeker Formation (Schalk, 1970). Resting unconformably on the Nosib Formation from which much of the detritus was derived, these rocks are composed of bedded conglomerates and pebbly quartzites. Many of the boulders and pebbles contain glacial striae attesting to the glacial origin of the deposit.

The final chapter in the development of the orogen was the Damara Orogeny which caused widespread deformation, metamorphism and intrusion. Movement was toward the cratons with the basement highs now acting as buttressing boundaries between the contrasting structural styles in the mobile Swakop Group and the openly-folded Otavi Group. Complex structural successions have been worked out in specific areas (e.g., Frets, 1969; Guj, 1970) and magmatic cooling events dated for the period 450-550 m.y. are well documented (Clifford, 1967; Burger and Coertze, 1973). This activity was a local expression of the Pan African event (Kennedy, 1964) which caused extensive orogenesis in widespread mobile belts, plus thermal resetting of radiometric clocks in older crystalline basement rocks throughout much of Africa, as well as Antarctica.

Gariep Group

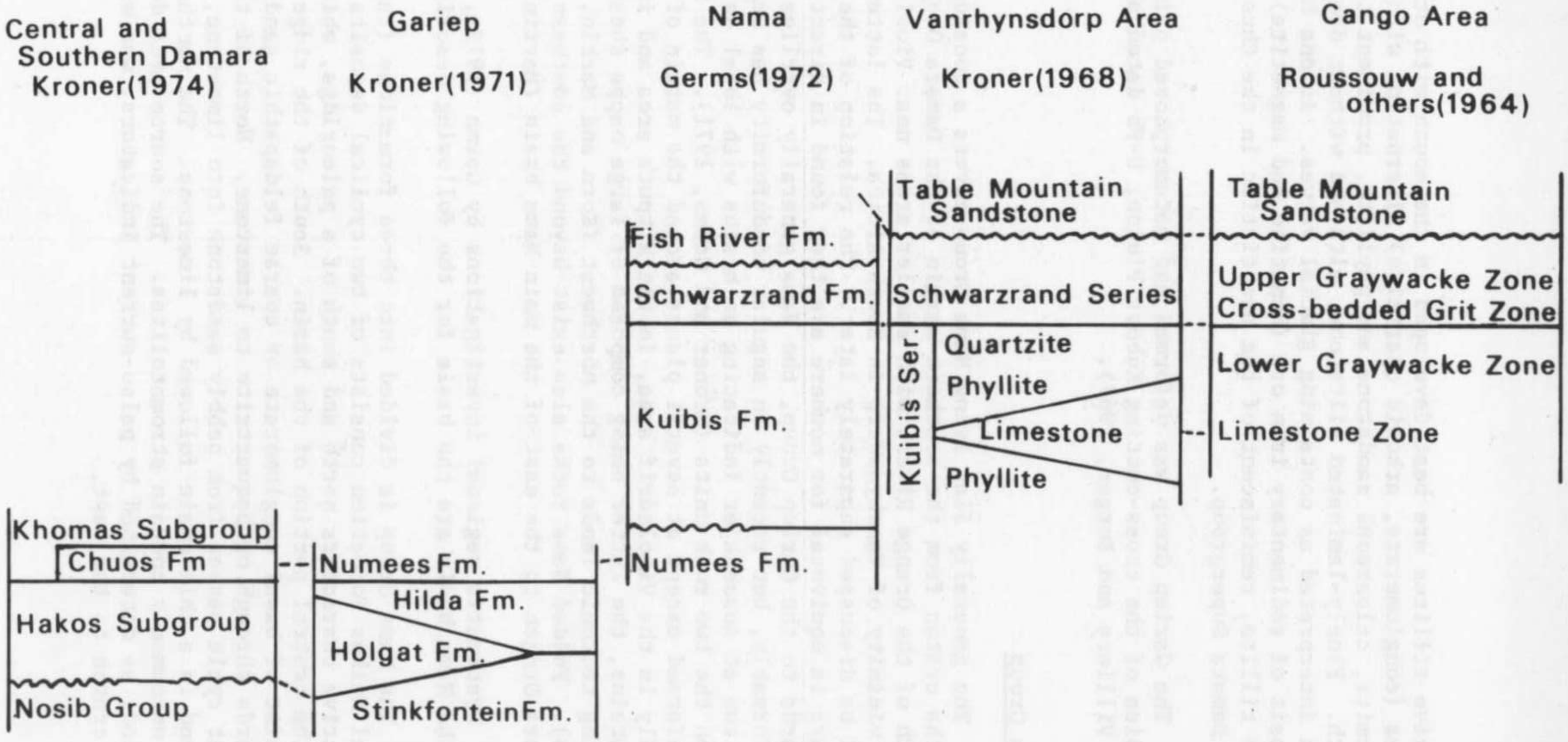
Deformed late Precambrian rocks appear farther south along the coast in a north-south trending belt between Luderitz in South West Africa and Kleinsee in South Africa. The most extensive work was done by de Villiers and Sohnge (1959) in the Richtersveld south of the Orange River. Their detailed descriptions are a landmark, although certain stratigraphic relations have since undergone reinterpretation. Numerous local names and uncertain relationships confuse the literature so I will follow the simplified subdivision put forth by Kröner (1971), whose final contributions on the Gariep Group are still in preparation (Table 8).

A geosynclinal cycle of sedimentation, here represented by the Gariep Group, is recognized. The initial deposition was that of the Stinkfontein Formation, a succession of quartzites, feldspathic quartzites and pelites, conglomerates, both basal and interformational, and interbeds of acid to intermediate lavas and pyroclastic rocks. The thickness of the formation is large and variable (possibly to 10 km, de Villers and Söhnge, 1959), and disappears completely in a northerly direction before reaching the Orange River. Martin (1965) suggests and Kröner (1971) affirms that the Stinkfontein is a correlative of the Nosib Formation farther to the north.

This is overlain to the east by the miogeosynclinal Hilda Formation and to the west by the eugeosynclinal Holgat Formation (Kröner, 1971). The Hilda Formation consists of carbonate and pelitic rocks whereas the Holgat Formation is principally graywacke, quartzite and schists with associated acid to basic volcanic rocks (Grootderm Series).

The Numees Formation follows. It fills two prominent north-northwest trending basins and several smaller ones (Martin, 1965). Fluvio-glacial and glaciomarine affinities with rapid facies changes are characteristic of this unit (Kröner and Rankama, 1972). Coarse

Table 8: Correlations between areas of the Damara Orogen, South and West Africa



massive tillites are best developed in the south with other clastic types (conglomerate, arkosic quartzite) alternating with limestone, dolomite, calcareous sandstone and phyllite, prominent toward the north. Finely-laminated siltstone, with and without drop stones, has been interpreted as containing glacial varves. At one horizon is a deposit of sedimentary iron ore (hematite and magnetite) intercalated with tillite, reminiscent of the association in the Chuos tillite of the Damara Supergroup.

The Gariep Group was deformed and metamorphosed prior to intrusion of the cross-cutting Kuboos Pluton, U-Pb dated at 550 ± 20 m.y. (de Villiers and Burger, 1967).

Nama Group

The generally flat lying Nama Group covers a considerable area of the craton from the southern margin of the Damara Orogen to just north of the Orange River, with smaller areas near Vioolsdrif and in the vicinity of Vanrhynsdorp in South Africa. The latter occurrence will be discussed separately later. The relation of the Nama to the Damara is equivocal for nowhere are they found in direct contact. With regards to the Gariep Group, the Nama generally overlies the Numees conformably, but recently an angular unconformity has been found between the two at Aussenkjer indicating an hiatus with local deformation between the two rock units (Kröner and Germs, 1971). The Nama Group is undeformed except at several places around the margin of its basin, namely in the Vioolsdrif area, in the Witputs area and in the Naukluft Mountains, the latter being composed of large nappe sheets shed from rising tectonic lands to the northwest (Korn and Martin, 1959; Martin, 1974). Folded Nama rocks also exist beyond the southern margin of the Damara Orogen to the east of the main Nama basin (Martin, 1965).

Systematic regional investigations by Germs (1972a, 1974) on much of the Nama basin are the basis for the following description.

The Nama Group is divided into three formations (Table 8). The basal Kuibus Formation consists of two cyclical deposits with distinctive characters north and south of a paleoridge, which existed in the central portion of the basin. South of the ridge both cycles consist of basal conglomerate or coarse feldspathic sandstone passing upwards through orthoquartzite to limestone. North of the ridge the first cycle passes from pebbly sandstone into limestone, while the second is a thin shale followed by limestone. The northern limestones not uncommonly contain stromatolites. The source area during this period, as determined by palso-current indicators, was mainly from the craton to the east.

During deposition of the overlying Schwarzrand Formation tectonic lands in the Damara Orogen to the north and northwest became an increasingly important sediment source, and the central ridge in the basin was eventually buried. The sediments consisting of quartzites and shales become more immature upwards and limestone members are important only in the southwestern portion of the basin. At times uplift within the Nama basin caused stream channels to be cut at levels within the formation. The Terminal Clastic Member lies upon a prominent erosion surface. Of note also is evidence in the form of grooved bedding planes and intraformational conglomerate for two periods of glacial activity in the time spanned by the Schwarzrand Formation.

The Fish River Formation was deposited upon another prominent erosion surface. Reddish immature sandstone, shale-pebble conglomerates and lack of limestone characterize these beds. Continued uplift in the Damara Orogen to the north is indicated from paleocurrent analysis and a western sediment source is recognized for the first time.

Correlation of the Nama with the Damara is a controversy which is still unresolved. The two most recent proposals have been by Kroner (1971) and Germs (1974). Both rely heavily on correlation of glacial events in the two areas. Kroner (1971) stated that the Numees Formation and Blaubeker Formation most certainly correspond, and further suggested that the Blaubeker Formation has affinity with the Nosib Formation and therefore cannot be related to the widespread intra-Damara Chuos Tillite. This then places the Chuos Tillite younger than the Numees of the Gariep Group and begs for a correlation with the more prominent glacial event recorded in the Schwarzrand Formation of the Nama Group.

Germs (1974) on the other hand favors a correspondence between the Numees and Chuos tillites, arguing that the Schwarzrand glacial event was minor when its deposits are compared to the much thicker and more extensive deposits of the Numees and Chuos. He also evokes fossil and tectonic evidence to suggest that the Nama must be younger than the Damara.

I tend to favor Germs' correlation for the reasons that he gives. But also because the tillite/magnetite-hematite association found in the Numees is distinctive among the intra-Damara tillites to the north. Kroner's argument that the Blaubeker Formation should be included with the Nosib also is without foundation for rocks of the Hakos and Khomas Subgroups do not occur over the southern margin of the Damara Orogen where the Blaubeker Formation is bounded by unconformaties above and below, between the Nama Group and the Nosib Group (Schalk, 1970).

Near the Orange River the Nama is intruded by the late syenite phase of the Bremen Igneous Complex dated at 550 ± 20 m.y. (U-Pb on zircons) and 506 ± 10 m.y. (Rb-Sr isochron) (Allsopp and others, 1974; Burger and Coertze, 1973).

Vanrhynsdorp Area

About 200 km south of the Orange River and 100 km north of the exposures of Malmesbury beds in the western Cape, one encounters the sedimentary and metamorphic rocks of the Vanrhynsdorp area, a group whose dual affinities with the Malmesbury of the Cape and the Nama of South West Africa have long been recognized. As in the rest of the Cape, work here was begun by the Geological Commission of the Cape of Good Hope (Rogers and Schwarz, 1900; Rogers, 1904, 1911, 1912). A three-fold division of Nieuwerust, Malmesbury and Ibiquas series was advocated by Rogers (1904), (Table 9). The Nieuwerust series was composed of varieties of arkose which onlapped the older granite and gneiss basement. The Malmesbury series was primarily slate and phyllite with a prominent zone of crystalline limestone in the middle and some bands of quartzite and feldspathic grit toward the top. The Ibiquas series, which began with a basal conglomerate and followed with slates, argillaceous sandstones with fine ripple marks and false bedding, and minor arkosic grits, was originally thought to rest with slight unconformity on the Malmesbury. Similarities of this sequence with the Nama System of South West Africa were regarded strong enough at that time to include the three series within it (Rogers, 1911, 1912).

TABLE 9. EVOLUTION OF STRATIGRAPHIC NOMENCLATURE, VANRHYNSDORP AREA

Rogers, 1904	Brink, 1950	Jansen, 1960	Kroner, 1968
Ibiquas	Schwarzkalk	Schwarzkalk	Schwarzrand
Malmesbury			Kuibis
Nieuwerust	Nuwerus(=Kuibis)	Kuibis	
		Malmesbury	

Since then several reports have concentrated on parts of the area (Lamont, 1947; Brink, 1950; Kroner, 1968; Buhrmann, 1969) and systematic quadrangle mapping has been carried out by the Geological Survey (Jansen, 1960; von Backstrom and others, 1960).

Following study of the area around Nuwerus, Brink (1950) demonstrated that the "basal" conglomerate of the Ibiquas was in fact an intraformational unit within the continuous sequence, and renamed the previous Malmesbury and Ibiquas series the Schwarzkalk Series correlative with the Nama unit of the same name in South West Africa. Also, the basal Nuwerus Series was correlated with the Kuibis Series of the Nama.

As a result of mapping by the Geological Survey (Jansen, 1960; von Backstrom and others, 1960) portions of the Namaqualand Gneiss Complex were interpreted as being derived from mobilized Malmesbury, of necessity older than this crystalline complex. Rogers' Ibiquas, now designated as Nama, was seen in sedimentary contact with the crystalline basement. Consequently all contacts between the Malmesbury and Nama were across postulated faults.

Kröner (1968, 1969), mapping selected small areas in detail, was able to refute the existence of these faults and returned to a conformable stratigraphy, now calling the Nuwerus and Malmesbury the Kuibis Formation and renaming the Ibiquas portion of Brink's (1950) Schwarzkalk, the Schwarzrand Series, after Martin's (1965) demonstration that the Schwarzkalk Series wedges out near the Orange River.

Bührmann (1969) studied an occurrence of rudities in the Kobe Valley and concluded that they were in part of glacial origin. The association with banded ironstones suggested correlation with the Numees glacials farther to the north. The overlying beds were given local formational names and a correlation with the basal Nama Group was suggested.

The rocks of the area were subsequently deformed (more intensely in the western portion) and metamorphosed to a low-grade. Later erosion was followed by deposition of the Table Mountain Group.

Pre-Cape Sediments

The closest lithologic similarities between the rocks of the Vanrhynsdorp area and the pre-Cape farther to the south are with those in the Cango Valley. Such a situation has long been recognized and several of the earlier authors suggested correlations (Hatch and Corstorphine, 1905; Rogers, 1905; Rogers and du Toit, 1909; MacIntyre, 1932). The Limestone Zone of the Cango, with its prominent limestone and phyllite and subsidiary lenses of arkosic grit and graywacke, bears a marked resemblance to the lower portion of the Kuibis around Vanrhynsdorp. Both are overlain by finer clastic rocks which are followed by a sudden influx of coarse material becoming reduced in grain-size up section (Cross-bedded Grit and Upper Graywacke Zones of the Cango and Schwarzrand Formation of the Vanrhynsdorp area).

If this correlation of the Vanrhynsdorp and Cango areas is correct there emerges a pattern of shallow-water and terrestrial basins situated marginally to the craton in a northwest to southeast arcing zone encompassing the Gamtoos, Cango and Worcester-Swellendam areas, the northeastern tectonic domain of the western Cape and the Vanrhynsdorp area, with the George area and central and southwestern tectonic domains of the western Cape located within a deep-water zone farther from the craton but parallel to its margin. This exterior zone became the locus of intrusion of most of the Cape Granites and is thought to have been the source of the coarse clastic sediments found in the upper strata of the interior zone, generated during the opening episode of the Damara Orogeny, which eventually encompassed rocks in all but the Gamtoos area.

Conclusions

If these correlations of rocks in the Vanrhynsdorp area with those to the south are correct, it is possible for the first time to relate events in the Cape to events in South West Africa by way of Kröner's (1968) correlations of the Vanrhynsdorp strata with the Nama farther to the north (Table 8). Although stratigraphic details vary greatly it appears possible that the increasingly immature sediments and northwesterly source of the Schwarzrand Formation in South West Africa record initial activity of the Damara Orogeny in that area as do the conglomeratic formations of the upper pre-Cape rocks in the Cape.

The timing of this event is not well established nor did it necessarily commence synchronously throughout the area. However, the fossil Cloudina found in the Kuibis and lower Schwarzrand Formations of the Nama is thought to be slightly older than 600 m.y., the base of the Cambrian (Germs, 1972a, 1972b). And as has been discussed previously (pp. 87-88) the 610 ± 20 m.y. date on the Cape Granite is suggestive of orogenic activity beginning at that time in the western Cape.

The Damara Orogeny is defined by a clustering of radiometric dates between 450 and 550 m.y. (Clifford, 1967) but this is not incompatible with initiation of the orogeny somewhat priot to 600 m.y., for age data, particularly K-Ar dates, indicate the times when isotopic systems become closed or homogenized. Therefore, many age determinations are more correctly related to the waning stages of orogenic activity when erosion has unroofed the deeper portions of the belt, rather than to the main paroxysm of an orogeny during which isotopic systems are most likely to be open (Armstrong, 1967).

Admittedly the proposed depositional event at the onset of the Damara Orogeny had decidedly different characteristics in different areas, but these may be explained by varying tectonic responses within the orogenic belt. In South West Africa the Nama basin formed on a relatively stable craton, with basement highs to the north and west acting as a boundary with the mobile orogenic belt. However, in the Cape south of the 1100 ± m.y. Namaqualand cratonic block on which the Nama has very limited development, there appears to be no such basement boundary between mobile belt and craton, and the basin which received the influx of coarse clastic sediments at the onset of orogenic activity was itself caught up in the expanding mobile belt in the Cape.

Part III: Comparison

COMPARISON: ROSS AND DAMARA OROGENS

Due to the apparent linear continuity of late Precambrian-early Paleozoic orogenic trends between the Transantarctic Mountains and southern Africa in reconstructions of Gondwanaland (see first chapter) a comparison of the geologic evolution of these two areas is warranted. When considering the Ross Orogen of Antarctica and the Damara Orogen of southern Africa in broadest form, one is at once struck by their similarity in isotopic ages and their differences in magmatism. In South West Africa dates cluster between 450 and 550 m.y. (Clifford, 1967) and in the Transantarctic Mountains most of the dates are between 450 and 520 m.y., (Grindley and McDougall, 1969), indicating that thermotectonic events were occurring simultaneously in these two portions of Gondwanaland and that their isotopic systems closed at the same time.

The magmatic development of the central Transantarctic Mountains was characterized by voluminous, pyroclastic, volcanic outpourings which were invaded by granitic intrusions of batholithic proportions, and as such resembled the tectonic province at the Pacific margin of the Americas. By contrast, in South West Africa the slight volcanism that occurred was of basic lavas, and the granodiorite covering a considerable area in the interior portion of the orogen appears to have been the product of granitization of the Nosib and Khomas Formations with only minor movement of magma as shown by local cross cutting relationships (Martin, 1965). Smith (1965) has determined that the two distinctive granite phases in the core of the belt are in general separated by the Chuos tillite and marble of the Hakos Subgroup. In South West Africa the relationships appear similar to those in the core of the Appalachians and Caldeonides with production of widespread ultrametamorphic products, but relatively minor actual plutonism (Rogers, 1970; Haller, 1971). Considering the Gondwana reconstructions in which the west coast of southern Africa is adjacent to South America and the Antarctic margin faces the Pacific Ocean, the contrast in tectonic style between the two areas may readily be explained by the collision of cratonic blocks in the former area and subduction of oceanic crust beneath a continental margin in the latter.

If this generalization is correct, a somewhat more detailed comparison should be instructive with regard to the nature of the transition between these two diverse tectonic regimes.

Dates on the Cape Granite and intruded Malmesbury Group (505-610 m.y.) are anomalously older by about 50 m.y. than the dates from South West Africa and Antarctica. If the preceeding arguments are correct that the initial phase of the Damara Orogeny occurred at about the same time throughout the African sector producing widespread clastic influxes, then the dates on the Cape Granite must be interpreted as indicating orogenic activity of approximately 50 m.y. shorter duration in the Cape than in South West Africa. The 610 ± 20 m.y. U-Pb date for formation of the Cape Granite (Burger and Coertze, 1973) and the suggested late Precambrian age for the Lower Kuibis Formation of the Nama Group (Germs, 1972a; 1972b) imply that this initial phase of the Damara Orogeny occurred during approximately the same period as the initial compressive phase of the Ross Orogen (Beardmore Orogeny) and that the unconformity on the graywacke-shale in the Nimrod Glacier area and Pensacola Mountains may be related to the initial erosion in the Damara Orogen. The data are too sparse to say whether the 20 m.y. difference between the Cape Granite date and the 630 ± m.y. dates from the Thiel and Horlick Mountains (Aaron and Ford, 1964; Faure and others, 1968) is significant.

Preceding this initial compressive phase, flysch sedimentation appears to have been the only sedimentary mode in Antarctica. In Africa, nearshore carbonate units (miogeosyncline) and deeperwater clastic units (eugeosyncline) accumulated concurrently. No older sediments are known in the Cape, but in central South West Africa the Khomas Subgroup overlies carbonate rocks of the Hakos Subgroup which in turn overlies with slight unconformity the arenaceous beds of the Nosib Group, except in the central portion of the belt where the contact is conformable (Smith, 1965). In a similar fashion in the Orange River area the Hilda and Holgat Formations of the Gariep Group overlie the arenacous Stinkfontein Formation (Kröner, 1971).

These clastic metasediments have been interpreted as representing discontinuous basin deposits derived from local remnant lands prior to full scale geosynclinal development (Martin, 1965). I would add that this relationship may indicate foundering of a belt between cratonic blocks, as opposed to deposits at a continental margin as developed in Antarctica. The carbonate facies-clastic facies division with lack of known underlying arenaceous beds in the Cape suggests that the sedimentary regime there was transitional between those in Antarctica and South West Africa.

The Damara Orogen apparently continued to be a positive area throughout its orogenic cycle following the initial uplift in the late Precambrian, except for molasse-type sediments of the Mulden Formation to the north. At some time during this period nappe sheets were shed from tectonic highlands to the northwest of the Nankluft Mountains into the Nama basin (Korn and Martin, 1959; Martin, 1974). In the Cape, sediments deposited during the initial tectonic activity were themselves caught up in folding as the mobile belt expanded, but there orogenic forces appear not to have been so persistent or severe as in central South West Africa, with recovery having been completed by the end of the Cambrian.

In the Ross Orogen the flysch eroded in the initial compressive phase was downwarped during the early to middle Cambrian and became the site of a back-arc basin adjacent to an active volcanic arc.

From the later Cambrian to middle Ordovician much of the inner portion of the Ross Orogen was invaded by intrusive granites and all of the preceding sedimentary and volcanic rocks were deformed, uplifted and eroded.

A tectonic contrast may be made between the Damara Orogen in South West Africa and the Ross Orogen in Antarctica. In South West Africa compression between cratonic blocks apparently caused the mobile belt to become a positive landmass from the initial tectonic unrest onwards and at the same time held down the core of the belt where ultrametamorphic effects produced in situ granitic rocks. In Antarctica the continental margin, compressed by subducting oceanic crust, behaved in a more flexible manner throughout the same period; that is, portions of the orogen between the magmatic arc and the craton were first folded and eroded, then downwarped to receive further sediments, and finally folded again and metamorphosed and intruded during the culminating phases of the orogeny. Age data indicate that tectonic forces relaxed at the same time in both South West Africa and Antarctica.

The Damara Orogen in the Cape appears to have aspects of both these two areas. Orogenic activity was not sustained as long as in South West Africa, and relatively minor amounts of granite were porduced, but neither was the effect as rapid or slight as in the Beardmore Orogeny, so that a second cycle of sedimentation did not become established as in Antarctica.

I would conjecture that the processes of "cratonization," which approached some physiochemical limit in the ultrametamorphic complex of South West Africa and in the batholithic complex of the central Transantarctic Mountains, were incomplete in the region transitional between these collisional and subductive regimes, and that this incomplete tectogenesis offered conditions conducive to the regional downwarp and succeeding folding spasm recorded in the Cape, Pensacola, and Ellsowrth Mountains during the latter half of the Paleozoic.

Post Script: Ellsworth Mountains

The Ellsowrth Mountains, which until now have not been mentioned, contain a stratigraphy which begins with a marble unit whose bottom is not exposed, overlain by a clastic unit that begins with conglomerates and grades upwards into increasingly finer-grained and shaley rocks. Following this conformably is a light-colored orthoquartzite with ubiquitous cross-beds (Craddock, 1969; Craddock and others, 1964). No granitic intrusions have been found in the succession. Upper Cambrian trilobites occur in a marble near the top of the pelitic clastic succession and Devonian brachiopods have been found near the top of the orthoquartzite (Webers, 1972). This stratigraphy is similar to that found in the Cape, and is the only occurrence in the region other than the Gamtoos area where an unconformity does not separate the basement rocks from the Beacon or Cape Supergroups. However, the Ellsworth Mountains are located on the exterior side of the Ross-Damara axis and have folds whose axes trend at right angles to those in the Pensacola Mountains 500 km to the southeast across the West Antarctic Ice Sheet.

In 1969 Schopf argued from stratigraphic relations in Upper Paleozoic rocks of the Ellsworth Mountains had been displaced during post-Jurassic continental breakup from a location at the margin of the East Antarctic craton somewhere north of the Pensacola Mountains. It seems unlikely that a basin could have remained tectonically calm on the exterior side of the Ross Orogen and received apparently continuous sedimentation during its considerable activity. The Gamtoos area in South Africa is in a more cratonward position in Gondwanaland than the mobile belt and probably not coincidentally the closest pre-Cape outcrop to Antarctica. These relationships incline me to agree with Schopf's (1969) Gondwana positioning of the Ellsowrth Mountains on the cratonic side of the Ross mobile belt in the notch left by the Weddell Sea when Antarctica is fitted to Africa. This requires adjustments in the paleogeographic maps previously published (Stump, 1973) which used a base map compatible with the concensus at the time (Hamilton, 1967; Frakes and Crowell, 1968; Elliot, 1972). This adjusted position gives a more rational distribution of facies than before and is compatible with the model developed in the Ross Orogen chapter.

APPENDICES

APPENDIX A: FIELD DESCRIPTIONS, QUEEN MAUD MOUNTAINS

This section systematically describes the geology of the field localities of non-intrusive basement rocks in the Queen Maud Mountains, Antarctica, visited during 1970-1971, and 1974-1975. Maps of each area accompany the text and are numbered sequentially beginning with those to the northwest. Figure 16 locates the areas.

Page

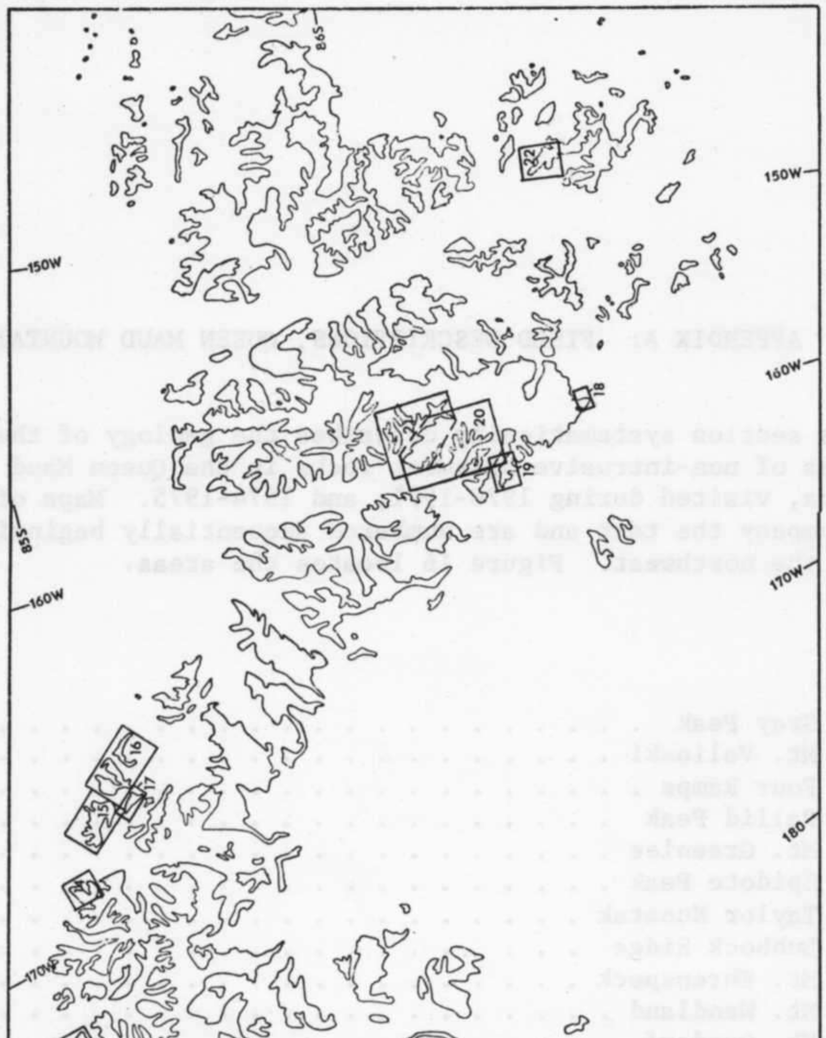
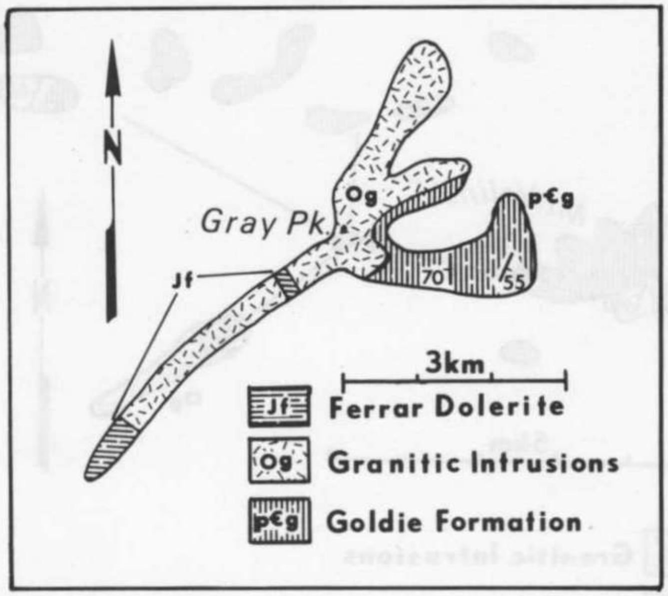


Figure 16. Locations of mapped areas, Queen Maud Mountains, Antarctica

1. Gray Peak, southeast ridge; 84°20'S, 173°58'E. Date: 14 December, 1970



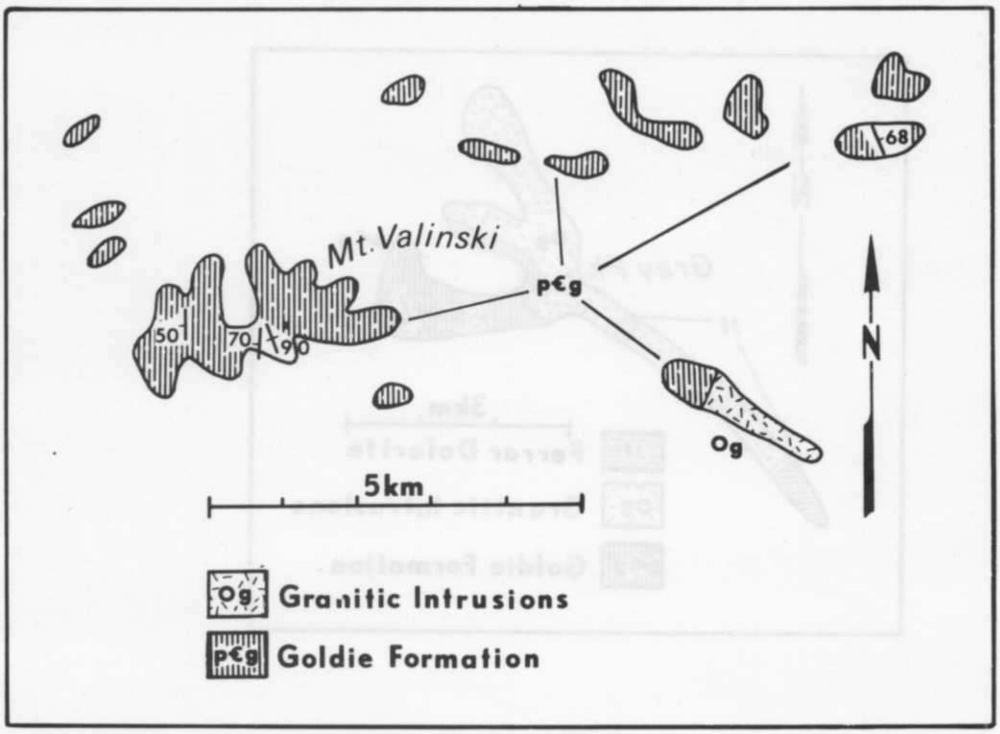
The Goldie Formation is intruded 2/3 of the way up the southeast ridge by the Hope Granite which underlies the peak as well as the northern and western spurs. Dikes of dolerite intrude the granite southeast of the peak.

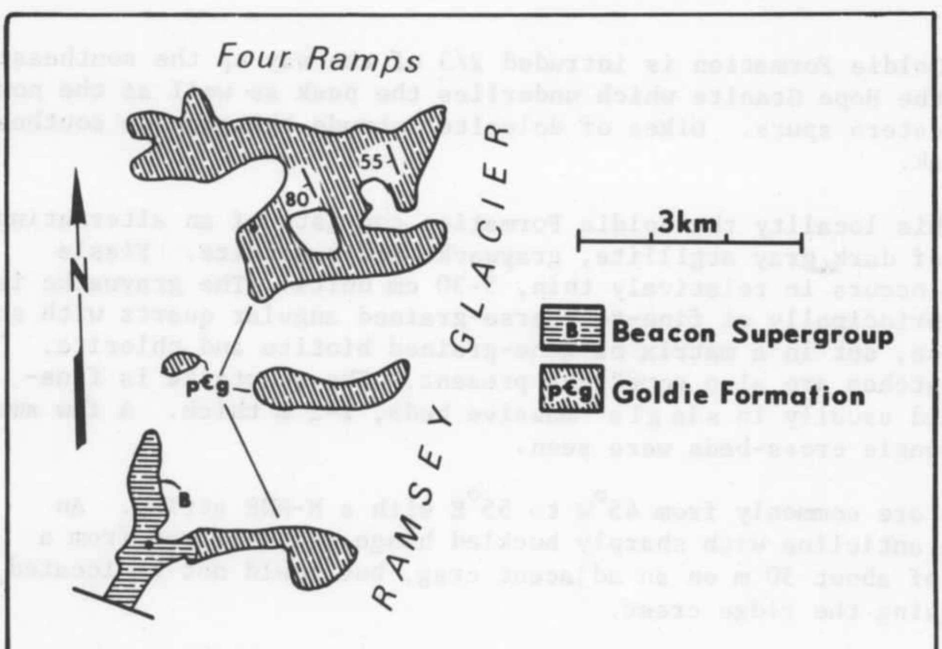
At this locality the Goldie Formation consists of an alternating sequence of dark gray argillite, graywacke and quartzite. Fissle argillite occurs in relatively thin, 5-30 cm units. The graywacke is composed principally of fine-to coarse-grained angular quartz with some plagioclase, set in a matrix of fine-grained biotite and chlorite. Calcite patches are also sometimes present. The quartzite is fine-grained and usually in single massive beds, 1-2 m thick. A few small, very low-angle cross-beds were seen.

Dips are commonly from 45°W to 55°E with a N-NNE strike. An isoclinal anticline with sharply buckled hinge was observed from a distance of about 50 m on an adjacent crag, but could not be located when crossing the ridge crest.

West side, upper Ramsey Glacier

2. Mt. Valinski and unnamed ridge north of Millington Glacier; 84°32'S, 177°20'E; date: 8 December, 1970.
3. Four Ramps; 84°42'S, 177°35'E; date: 29 November, 1970.





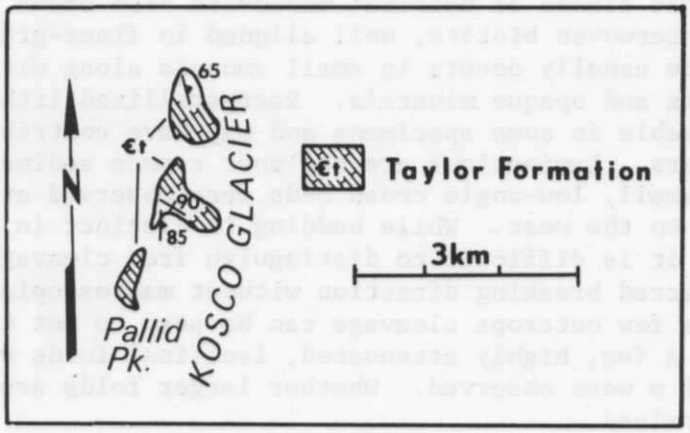
Goldie Formation crops out at both of the above locations. The rocks strike N-NW and dip from 50°W to vertical, being overturned at Mt. Valinski.

At Mt. Valinski the rocks consist of alternating dark gray, thin-bedded to fissile argillite and medium to thick-bedded graywacke, with occasional beds of massive, fine-grained quartzite. The graywacke is fine- to medium-grained with angular quartz and lesser amounts of plagioclase. Sparse flakes of detrital muscovite also occur. The matrix contains interwoven biotite, well aligned in finer-grained specimens. Calcite usually occurs in small amounts along with recrystallized quartz and opaque minerals. Recrystallized lithic fragments are recognizable in some specimens and may have contributed to the matrix in others. Laminations are the most common sedimentary feature although small, low-angle cross beds were observed at several places, with tops to the west. While bedding is distinct in some places, in others it is difficult to distinguish from cleavage, which appears as a preferred breaking direction without macroscopic mica growth. Also in a few outcrops cleavage can be seen to cut the bedding at a high angle. A few, highly attenuated, isoclinal folds with amplitudes up to 1 m were observed. Whether larger folds are present could not be determined.

A hard, dark gray, fine- to medium-grained, calcite-bearing quartzite underlies the ridge north of Millington Glacier.

At Four Ramps the rock compositions and bedding relations are similar to those at Mt. Valinski, but a somewhat higher grade of metamorphism occurs. The rocks have a green coloration which is more pronounced toward the west. Biotite, sericite and muscovite are larger in size forming a crude foliation in some rocks. In one sample, a phyllite, fine-grained biotite has grown in a conjugate pattern. Vein quartz is common throughout much of the finer-grained portions of the rock.

4. Pallid Peak, west side of Kosco Glacier; 84°36'S, 178°39'W; date: 3 December, 1970



Outcrops occur on the three small peaks at the northern end of the ridge. A very coarsely crystalline, snow white marble, probably correlative with the Henson Marble, makes up most of the exposure.

The rocks is massive and non-bedded most places, but some faint, thick (1-3 m) beds can be discerned. Fe-oxide staining is prevalent in parts. A foliation is roughly developed through the calcite grains and usually cuts bedding where both can be seen. Microscopically, many of the grains are twinned. Thin (1-3 cm) beds, some laminated, of fine- to medium-grained, light gray calcite occur on the northern small peak. This bedding is quite straight for the most part but can be traced into a 30 m zone of breccia, individual blocks of which average less than 0.5 m, but are up to 2 m in length. Filamentous forms separate the blocks, attesting to the extreme mobility of this rock at the time of deformation. Mesoscopic folds were observed in some cherty and quartzitic beds on the southern peak. Patches of tan, columnar actinolite, up to 5 mm in length, are sometimes associated with the siliceous layers.

Northern Shackleton Glacier area; 84°45'S, 176°15'W

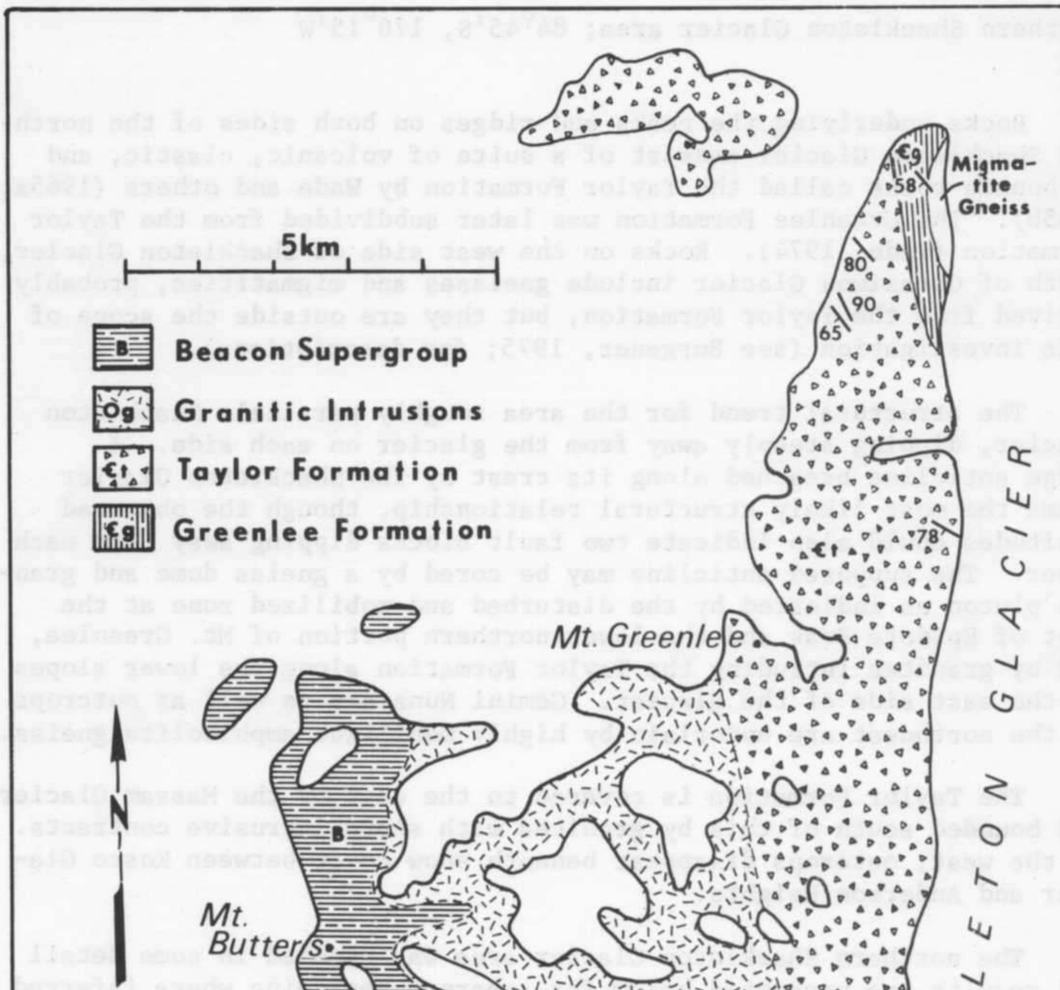
Rocks underlying the peaks and ridges on both sides of the northern Shackleton Glacier consist of a suite of volcanic, clastic, and carbonate rocks called the Taylor Formation by Wade and others (1965a; 1965b). The Greenlee Formation was later subdivided from the Taylor Formation (Wade, 1974). Rocks on the west side of Shackleton Glacier, north of Gerasimou Glacier include gneisses and migmatities, probably derived from the Taylor Formation, but they are outside the scope of this investigation (see Burgener, 1975; for descriptions).

The structural trend for the area roughly parallels Shackleton Glacier, dipping steeply away from the glacier on each side. A large anticline breached along its crest by the Shackleton Glacier seems the most likely structural relationship, though the observed attitudes could also indicate two fault blocks dipping away from each other. The supposed anticline may be cored by a gneiss dome and granite pluton as indicated by the disturbed and mobilized zone at the foot of Epidote Peak and the lower northern portion of Mt. Greenlee, and by granites intruding the Taylor Formation along the lower slopes on the east side of the glacier. Gemini Nunataks as well as outcrops to the northwest are underlain by highly mobilized amphibolite gneiss.

The Taylor Formation is covered to the east by the Massam Glacier and bounded south of this by granites with sharp intrusive contracts. On the west, outcrops disappear beneath snow cover between Kosco Glacier and Anderson Heights.

The northern Shackleton Glacier area was studied in some detail and results are presented below for subareas beginning where inferred older strata occur.

5. Mt. Greenlee; 84°50'S, 177°00'W; dates: 15, 20, 22 and 27 November, 1970.



Mt. Greenlee is underlain by sedimentary and volcanic rocks which strike NNW to NW and dip steeply toward the SW. The southwestern portions of the mountain are underlain by basement granite and sedimentary rock of the Beacon Supergroup. Outcrops at glacier level and on the northern ridge crest were visited.

On the northern end of Mt. Greenlee at glacier level is a zone of highly disturbed and mobilized rock. Within it are layers and chunks of dark biotite and hornblende gneiss which were presumably derived from the overlying metasediments. A crude stratigraphy is retained in the layered portions, while elsewhere the blocks may be aligned or disoriented. The gneiss is surrounded by white fine-grained hornblende-bearing granitic rock, which has partially resorbed some of it, and reacted little or not at all with the rest. The entire mass is cross cut by pegmatite veins composed of quartz, plagioclase, K-feldspar, biotite and epidote. This exposure appears to be the contact zone at the roof of a gneiss dome or granite pluton at depth.

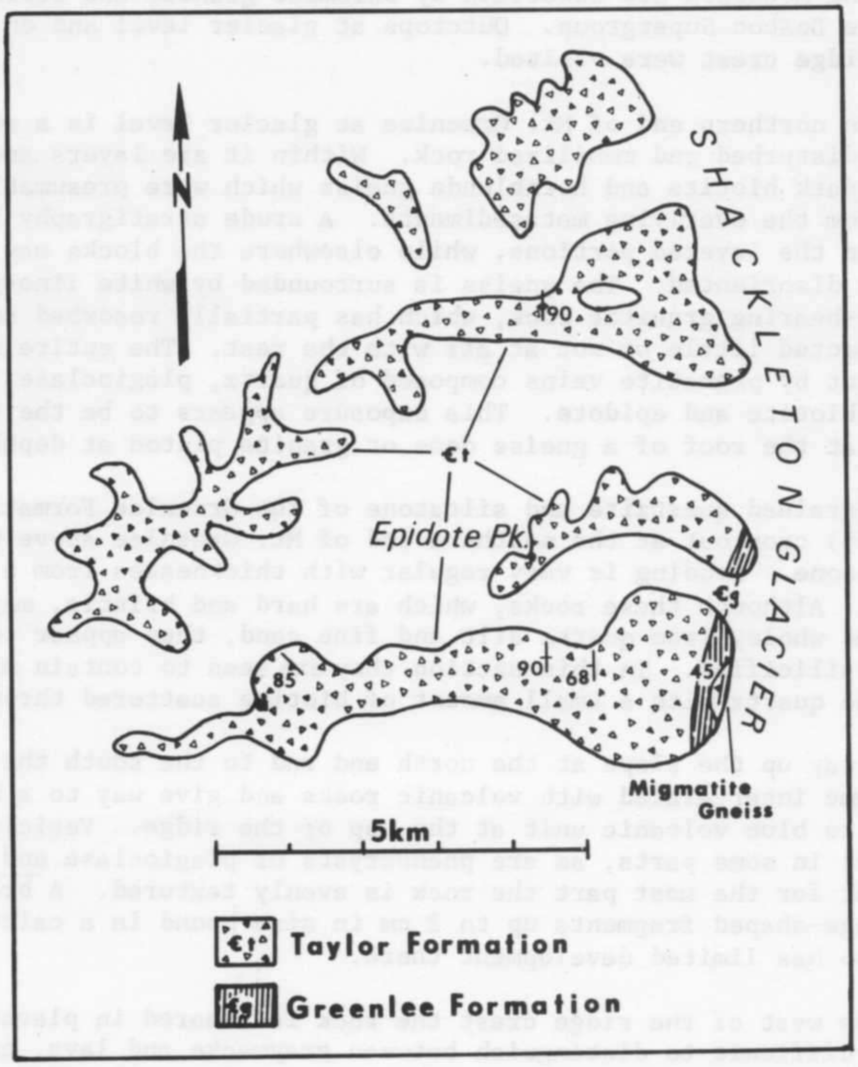
Fine-grained quartzite and siltstone of the Greenlee Formation (Wade, 1974) crop out at the northern end of Mt. Greenlee above the disturbed zone. Bedding is very regular with thicknesses from a few cm to 1 m. Although these rocks, which are hard and brittle, may have been formed wholey from quartz silt and fine sand, they appear to have been silicified. In thin section they are seen to contain microcrystalline quartz with a small amount of biotite scattered throughout.

Part way up the slope at the north end and to the south the sediments become intercalated with volcanic rocks and give way to a massive dark gray to blue volcanic unit at the top of the ridge. Vesicles are present in some parts, as are phenocrysts of plagioclase and rarely quartz, but for the most part the rock is evenly textured. A breccia with lozenge-shaped fragments up to 2 cm in size bound in a calcite matrix also has limited development there.

To the west of the ridge crest the rock is sheared in places, and it is difficult to distinguish between graywacke and lava, each of which can be identified when unsheared. Argillite, limestone and "chert" are also encountered in this area. Farther to the west the sediments again give way to dark, massive, volcanic rock.

In a few places in the sedimentary rocks a cleavage is developed striking parallel to bedding and with a near vertical dip.

6. Epidote Peak; 84°46'S, 176°56'W; Dates: 20, 21, 23 November, 1970.



The rocks at Epidote Peak are similar to those found at the northern end of Mt. Greenlee. On the east flank at glacier level there is a zone of highly mobilized rock, above which is a sequence of evenly bedded (5 cm-1 m) argillite and fine-grained quartzite. Laminations are common in many beds.

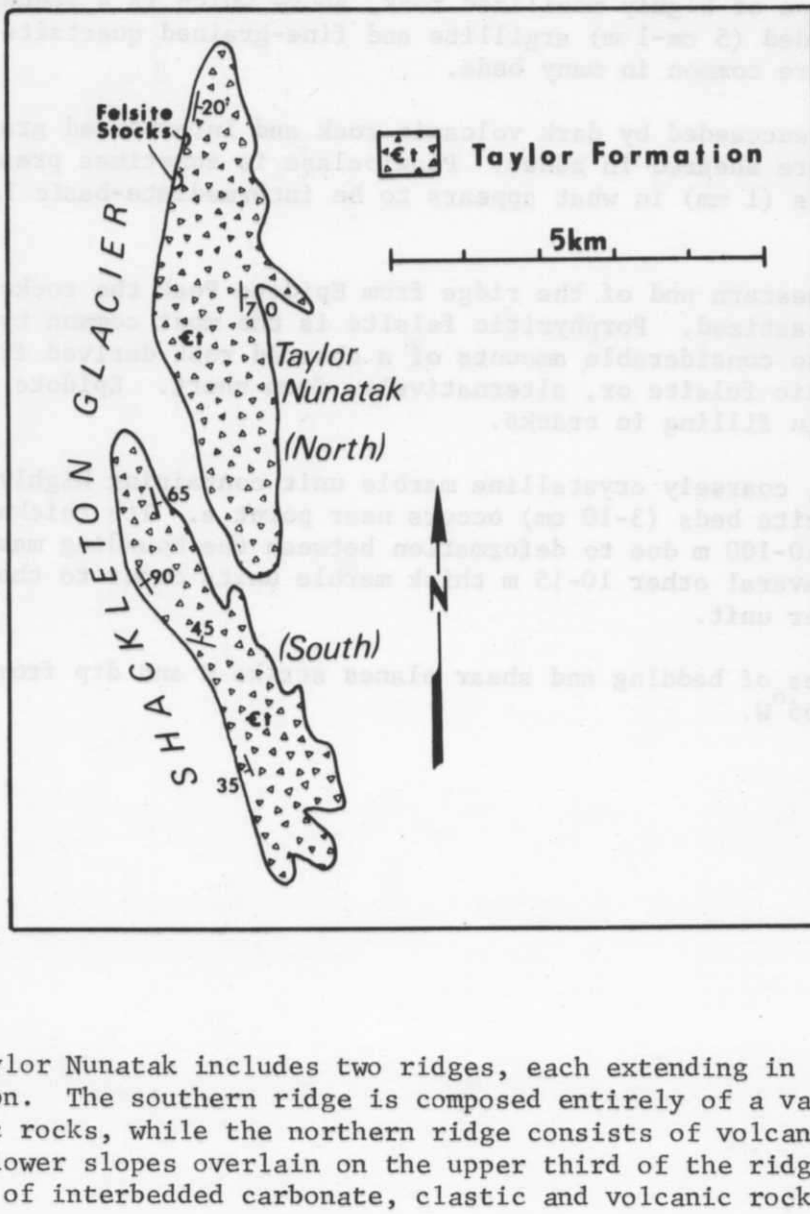
This is succeeded by dark volcanic rock and interbedded graywacke which are sheared in zones. Plagioclase is sometimes present as phenocrysts (1 mm) in what appears to be intermediate-basic lava flows.

At the western end of the ridge from Epidote Peak the rocks are highly cataclastized. Porphyritic felsite is the most common type but there are also considerable amounts of a sheared rock derived from non-porphyritic felsite or, alternatively, from chert. Epidote is common as vein filling in cracks.

A white, coarsely crystalline marble unit containing highly folded quartzite beds (3-10 cm) occurs near point a. Its thickness varies from 10-100 m due to deformation between the bounding massive felsites. Several other 10-15 m thick marble units occur to the east of the thicker unit.

Attitudes of bedding and shear planes strike N and dip from near vertical to 65°W.

7. Taylor Nunatak; $84^{\circ}54'S$, $176^{\circ}10'W$; dates: 11, 12, November, 1975; 19, December, 1975.



Taylor Nunatak includes two ridges, each extending in a N-S direction. The southern ridge is composed entirely of a variety of volcanic rocks, while the northern ridge consists of volcanic rocks on its lower slopes overlain on the upper third of the ridge by a section of interbedded carbonate, clastic and volcanic rocks.

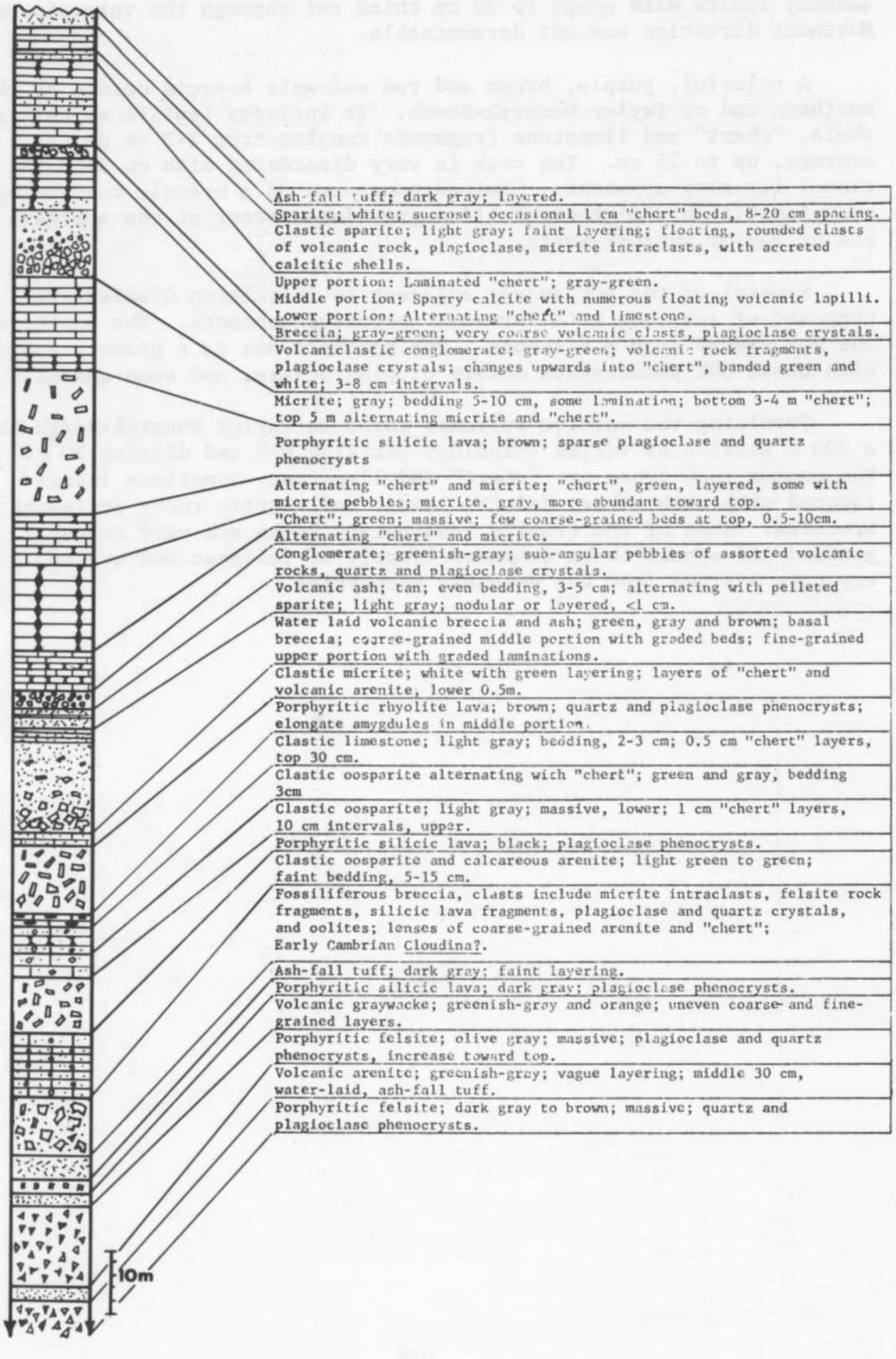
The predominant rock type on both ridges is a dark gray, massive felsite with phenocrysts of rounded quartz and euhedral plagioclase. Faint layering, perhaps a flow-banding texture, was observed at a few places in this rock, which otherwise is devoid of emplacement features. Attitudes are quite variable at Taylor Nunatak-South probably indicating some folding there. Toward the southern end several faults with gouge to 30 cm thick cut through the volcanic rock. Movement direction was not determinable.

A colorful, purple, brown and red volcanic breccia occurs at the northern end of Taylor Nunatak-South. It includes lapilli as well as shale, "chert" and limestone fragments ranging from 2-3 cm on the average, up to 15 cm. The rock is very disordered with no depositional layering apparent. Limited exposures of a breccia containing only lapilli in a microcrystalline groundmass occur at the northern end of Taylor Nunatak-North.

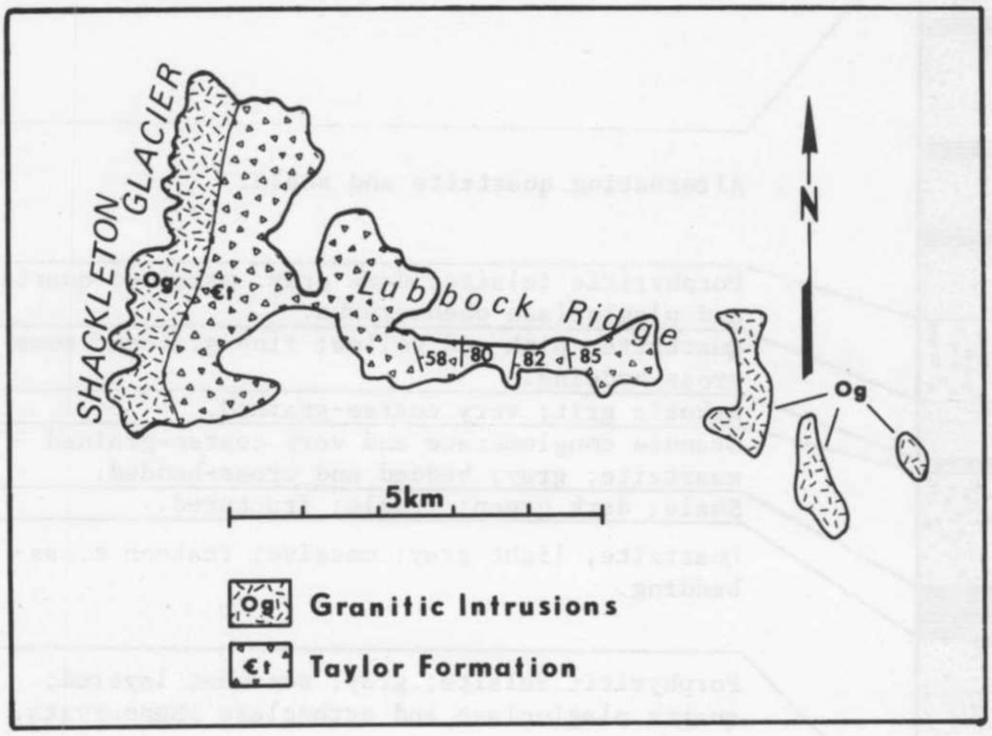
Several of the buttresses adjacent to Shackleton Glacier are composed of felsites with vertical intrusive contacts. The southern one was viewed from a distance. The northern one is a green porphyry with about 50% phenocrysts mostly of plagioclase, and some quartz.

Overlying the massive volcanic rocks at Taylor Nunatak-North is a 200 m section of varied lithology striking N-S and dipping 70°E. The common rock types are "chert" and limestone, sometimes interlayered with each other, felsite lavas, and clastic rocks including breccias. Some of the clastic sedimentary rocks are very coarse-grained and appear to be composed largely of feldspar and quartz crystals derived from the volcanic porphyries.

Figure 17. Stratigraphic Section, Taylor Formation - Taylor Nunatak



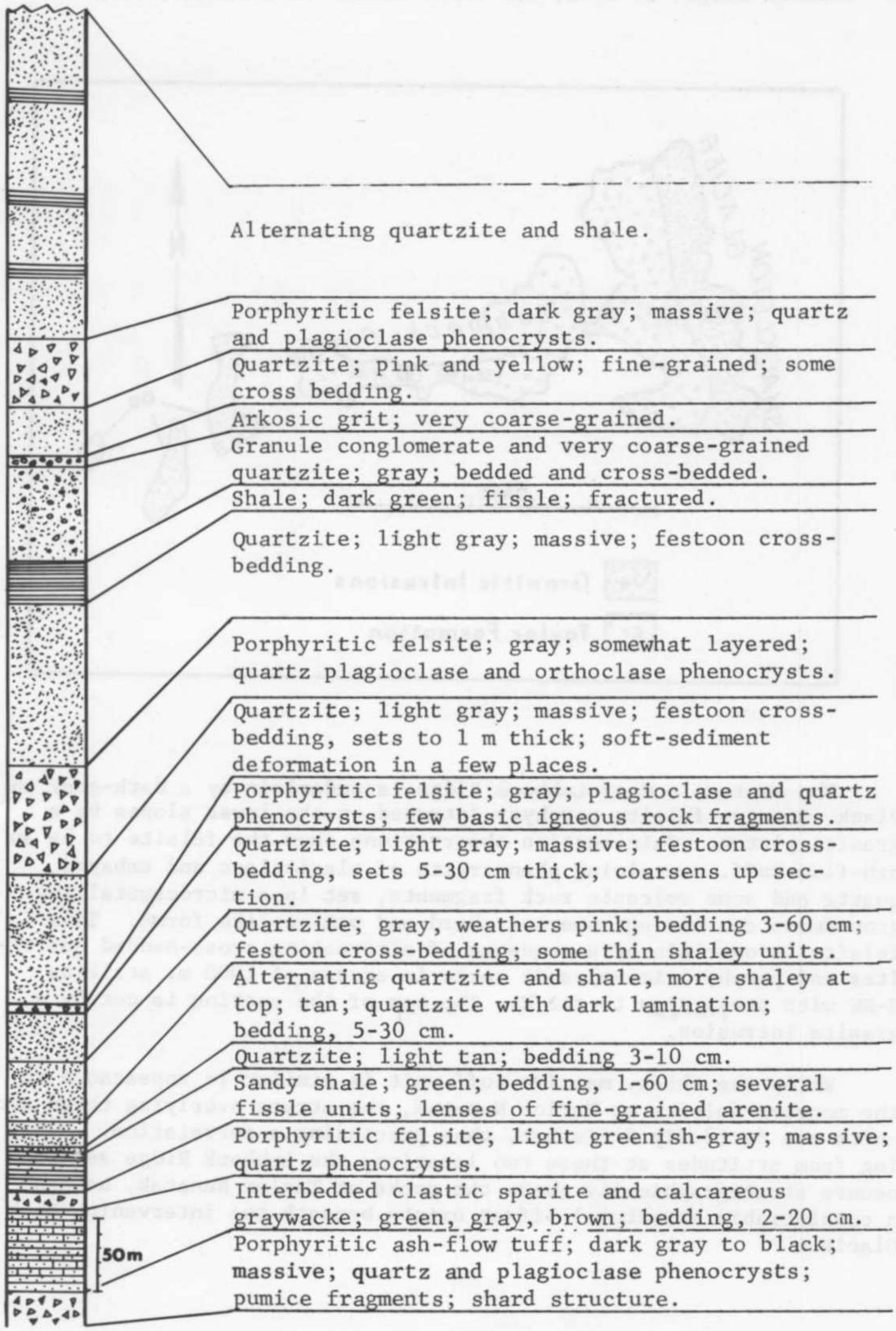
8. Lubbock Ridge; 84°50'S, 175°50'W; date: 12 November, 1975



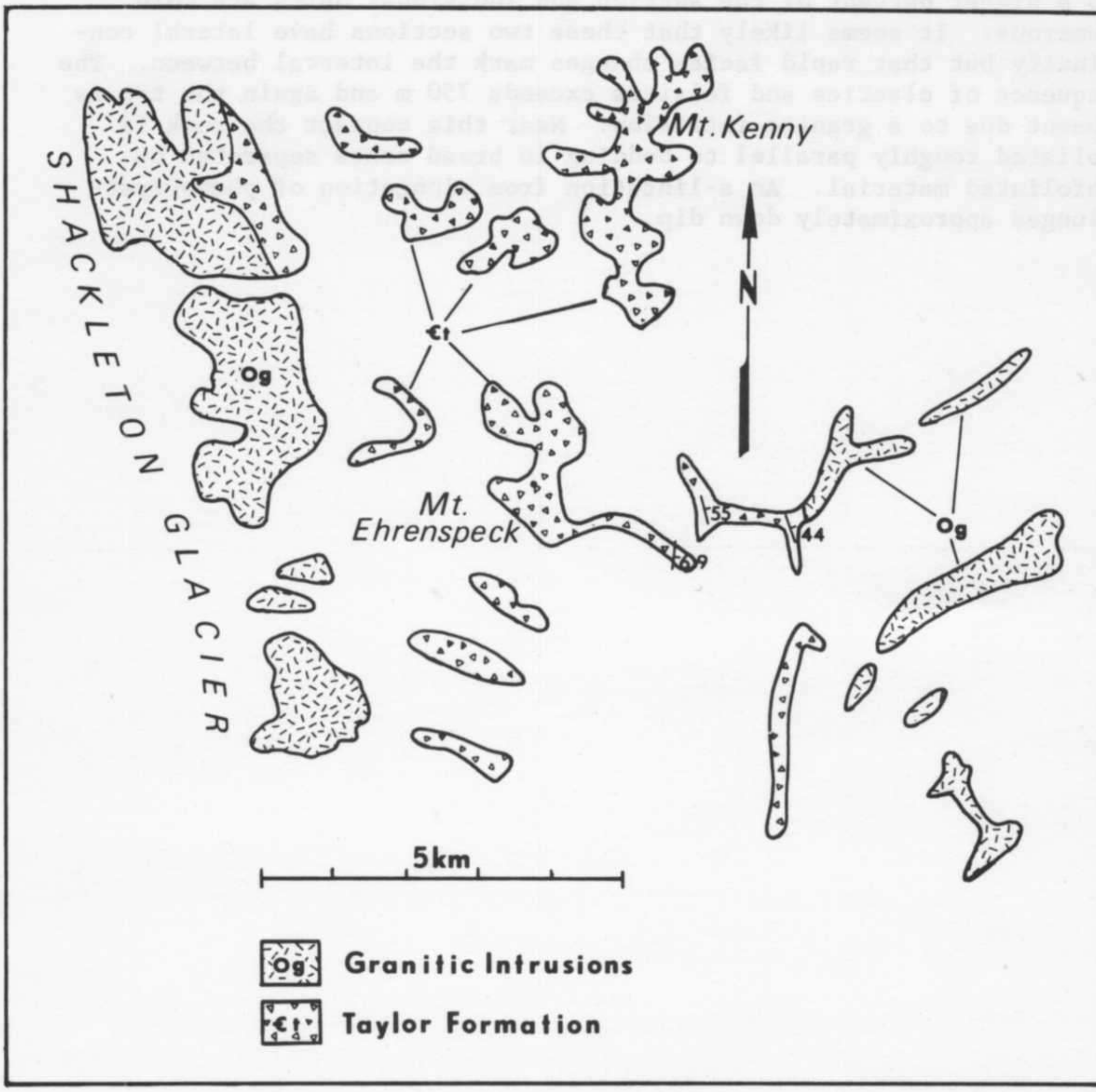
The western end of Lubbock Ridge is underlain by a dark-gray to black, massive felsite porphyry intruded on the lower slopes by a granite pluton. Thin-section observations show the felsite to be an ash-flow tuff, containing phenocrysts of plagioclase and embayed quartz and some volcanic rock fragments, set in a microcrystalline groundmass containing numerous shard and pumice-like forms. This felsite is overlain by a sequence of alternating cross-bedded quartzites and porphyritic volcanic rocks in excess of 1000 m, striking N-NW with steep dips to the E. The top of the section is cut by a granite intrusion.

While the thick, massive tuff unit is similar in appearance to the massive felsite at Taylor Nunatak, the strata overlying these two units are decidedly different, thus precluding a correlation. Judging from attitudes at these two locations the Lubbock Ridge section occurs stratigraphically above the rocks at Taylor Nunatak, unless a considerable structural offset exists beneath the intervening Dick Glacier.

Figure 18. Stratigaraphic Section, Taylor Formation - Lubbock Ridge

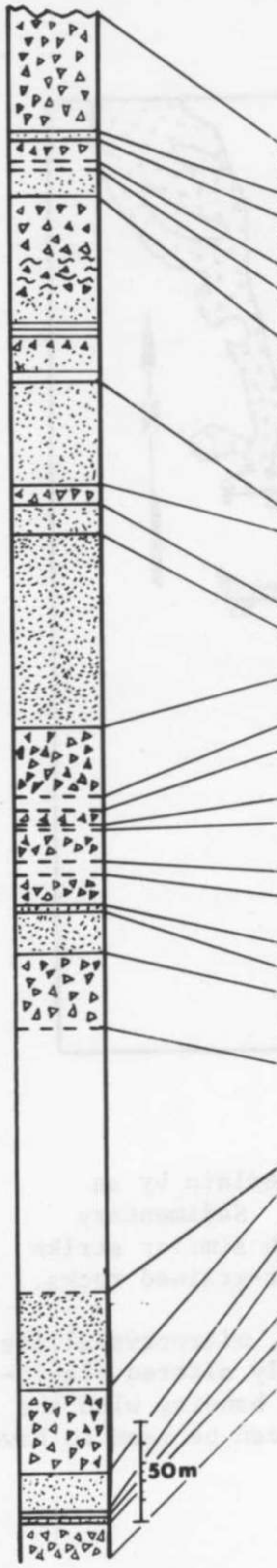


9. Mt. Ehrenspeck; 84°46'S, 175°30'W; date: 17 November, 1970



On the narrow ridge crests extending east from Mt. Ehrenspeck there occurs a sequence of alternating cross-bedded quartzites and porphyritic felsites, striking N-NW with 45°-70°E dips. Underlying the peak and the spurs to the west, as seen from helicopter reconnaissance, is a dark, massive volcanic rock, intruded by granite high on the oversteepened slopes above Shackleton Glacier. The association is analogous to the one on Lubbock Ridge; however, here felsite makes up a higher percent of the section and individual units are more numerous. It seems likely that these two sections have lateral continuity but that rapid facies changes mark the interval between. The sequence of clastics and felsites exceeds 750 m and again the top is absent due to a granite intrusion. Near this contact the rock is foliated roughly parallel to bedding in broad zones separated by unfoliated material. An a-lineation from elongation of phenocrysts plunges approximately down dip.

Figure 19. Stratigraphic Section, Taylor Formation-
Mt. Ehrenspeck



Porphyritic rhyolite; gray; foliated; quartz and plagioclase phenocrysts.

Dark, fine-grained quartzite and fissle shale.

Porphyritic felsite; gray; foliated, quartz and plagioclase phenocrysts.

-- Snow cover--

Gray to white, foliated, gritty quartzite; contains augens of quartz grains and fine-grained quartzite.

Alternating felsite and sediment.
Upper portion: Porphyritic felsite; dark gray; massive with several foliated zones; quartz phenocrysts.
Middle portion: Porphyritic felsite; gray; mylonitized; quartz and plagioclase phenocrysts.
Lower portion: Alternating porphyritic felsite, quartzite and argillite.

Quartzite; tan to pink; massive.

Porphyritic rhyolite; brown with orange splotches; quartz, plagioclase and orthoclase phenocrysts, broken chips somewhat aligned.

Quartzite; tan and pink; massive.

Quartzite; gray and pink; cross-bedding; very coarse unit at bottom.

Porphyritic felsite; pink; massive; quartz, plagioclase and orthoclase phenocrysts.

--Snow cover--

Porphyritic felsite; pink; massive; quartz, plagioclase and orthoclase phenocrysts.

--Snow cover--

Porphyritic rhyolite; pink; massive; quartz, plagioclase and orthoclase phenocrysts.

--Snow cover--

Porphyritic felsite; pink; massive; quartz, plagioclase and orthoclase phenocrysts.

Quartzite; gray; very coarse-grained.

Quartzite; white; cross-bedding.

Porphyritic rhyolite; gray; massive; quartz, plagioclase and orthoclase phenocrysts; altered basic igneous rock fragments(?).

--Snow cover--

Quartzite; white to gray; cross-bedding.

Porphyritic rhyolite; gray; massive; quartz, plagioclase and orthoclase phenocrysts.

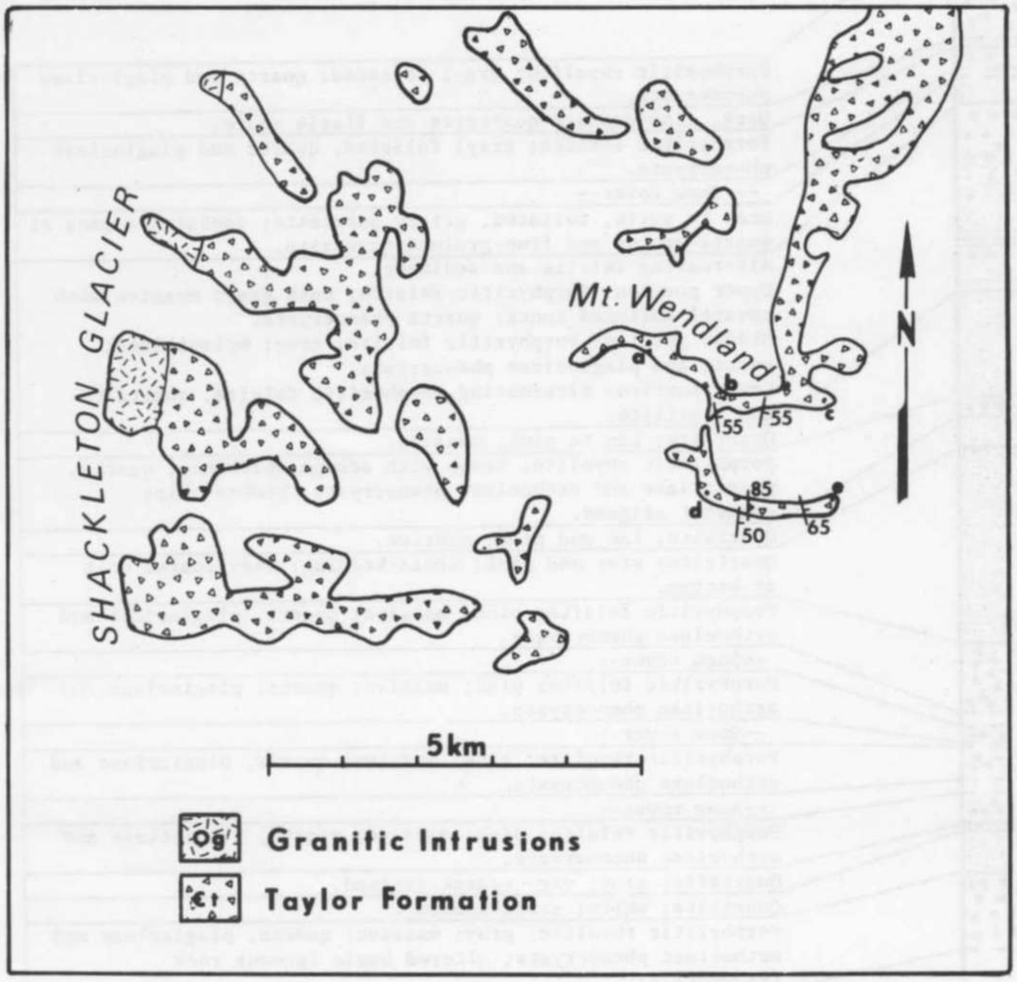
Quartzite; white; massive.

Porphyritic felsite; reddish brown; quartz and feldspar phenocrysts.

Quartzite; white; bedding, 3-30 cm; indistinct cross-bedding.

Porphyritic felsite; reddish brown; massive; quartz and feldspar phenocrysts.

10. Mt. Wendland; 84°42'S, 175°20'W; date: 18 November, 1970



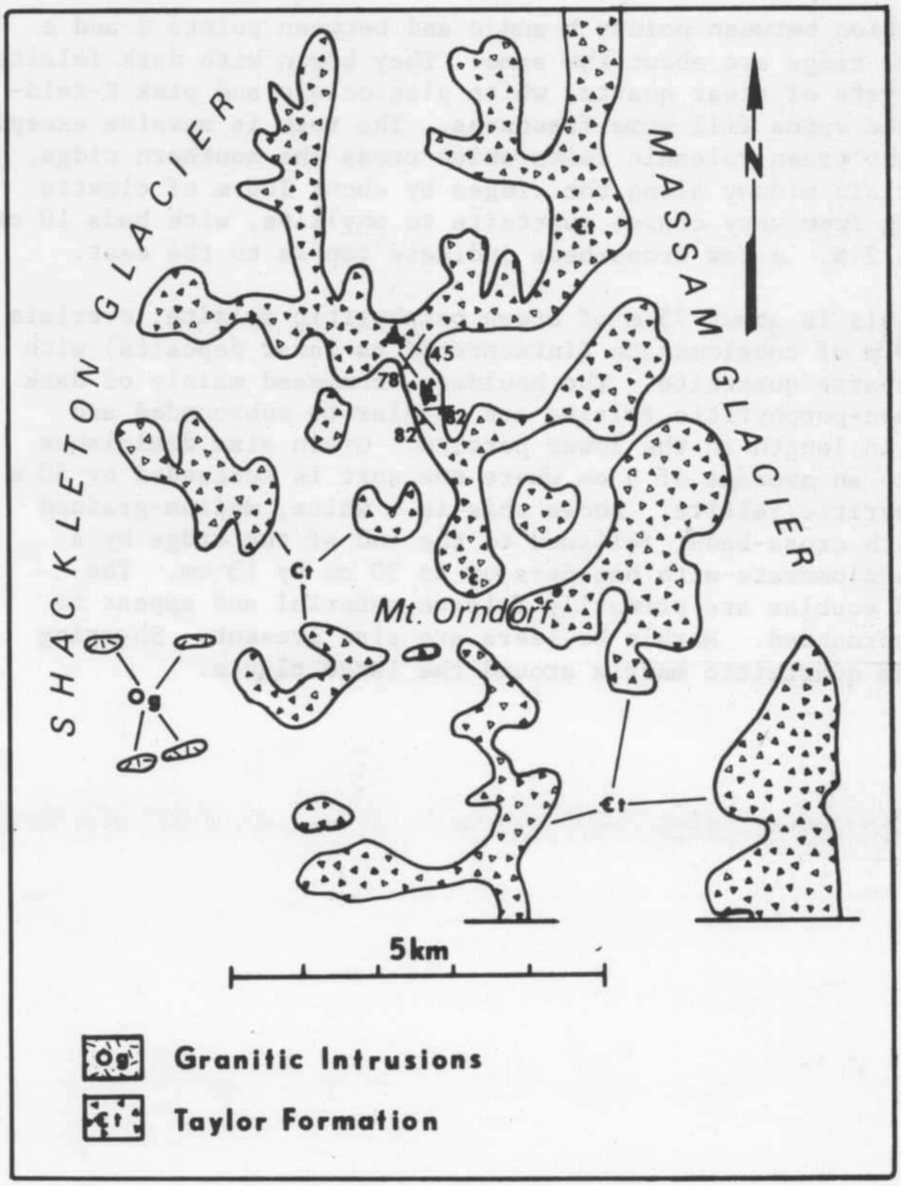
The ridge crests south of Mt. Wendland are underlain by an alternating sequence of volcanic and clastic rocks. Sedimentary rocks strike N-NW and dip 50°-75°E. A cleavage with similar strike and steeper easterly dip occurs in some of the finer-grained rocks.

Between points a and b the rock is mostly dark, microcrystalline felsite with phenocrysts of embayed quartz and highly altered plagioclase. In thin section one sample exhibits a dusky banding with the appearance of welded shard structure. Spherulites can be seen to have grown across the banding, in some cases offsetting it slightly. Some light colored "chert" also occurs in this interval. Near point b is a breccia with faintly layered blocks of darp porphyritic felsite up to 30 cm in size, surrounded by light material composed largely of quartz grains (2-3 mm), probably phenocrysts.

The section between points b and c and between points d and e on a parallel ridge are about the same. They begin with dark felsite with phenocrysts of clear quartz, white plagioclase and pink K-feldspar. Epidote veins fill some fractures. The rock is massive except for four light green volcanic units which cross the southern ridge. This is overlain midway along the ridges by about 100 m of clastic rocks ranging from very coarse quartzite to phyllite, with beds 10 cm to more than 2 m. A few cross-beds indicate top is to the east.

Above this is about 75 m of brown porphyritic felsite, overlain by about 100 m of conglomerate (interpreted as lahar deposits) with intermixed coarse quartzite. The boulders, composed mainly of dark "chert" or non-porphyritic felsite are angular to subrounded and up to 30 cm in length in the lower portion. Grain size diminishes up-section to an average of 3 cm where the unit is succeeded by 50 m of tan porphyritic felsite. Above this is a white, medium-grained quartzite with cross-beds, followed to the end of the ridge by a stretched conglomerate with boulders up to 30 cm by 15 cm. The boulders and cobbles are primarily felsite material and appear to have been subrounded. Marble boulders are also present. Shearing occurs in the quartzitic matrix around the large clasts.

11. Mt. Orndorf; 84°37'S, 175°32'W; date: 5 December, 1970.



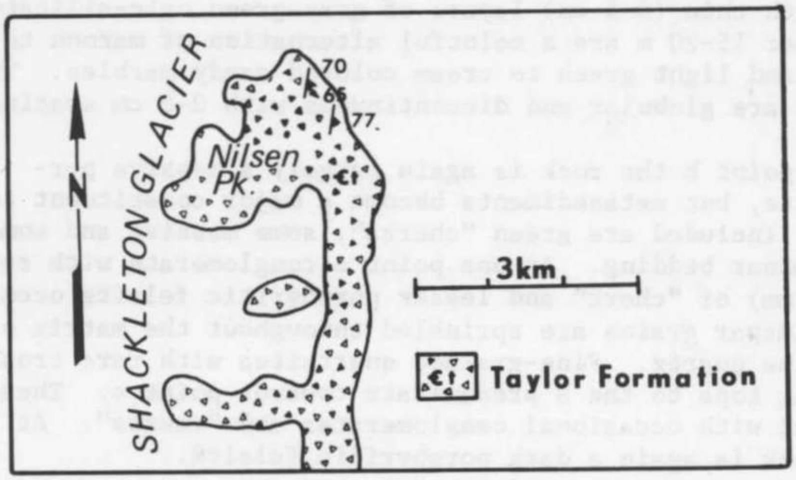
Rocks north of Mt. Orndorf include porphyritic felsite, "chert", cross-bedded quartzite and marble. Between points a and b the rock is a massive green to gray, microcrystalline felsite with rounded quartz phenocrysts, and some parts euhedral plagioclase crystals. Also incorporated are fragments of darker volcanic rock, some containing phenocrysts.

A 30-40 m thick marble unit occurs at point b. The lower 15-20 m are white, bedded (< 20 cm), fine- to medium-grained marble, interbedded with thin (3-5 cm) layers of gray-green calc-silicate rock. The upper 15-20 m are a colorful alternation of maroon to pink cherty layers and light green to cream colored sandy marbles. The "chert" layers are globular and discontinuous with 2-8 cm spacing.

South of point b the rock is again largely a massive porphyritic felsite, but metasediments become a major constituent toward point c. Included are green "cherts", some massive and some with faint laminar bedding. At one point a conglomerate with rounded pebbles (to 8 mm) of "chert" and lesser porphyritic felsite occurs. Quartz and feldspar grains are sprinkled throughout the matrix of microcrystalline quartz. Fine-grained quartzites with rare cross-beds indicating tops to the S predominate towards point c. These are interbedded with occasional conglomerates and "cherts". At point c the rock is again a dark porphyritic felsite.

Bedding strikes from NE to E with dips 45^{o}-85^{o}S. A prominent cleavage is developed in some of the finer-grained layers with NW strike and dips about 80^{o}S. While the cleavage is consistent with measurements recorded to the south along the east side of Shackleton Glacier, the bedding has a decidedly different orientation. Time did not permit visiting intervening outcrops so the structural situation there is undetermined.

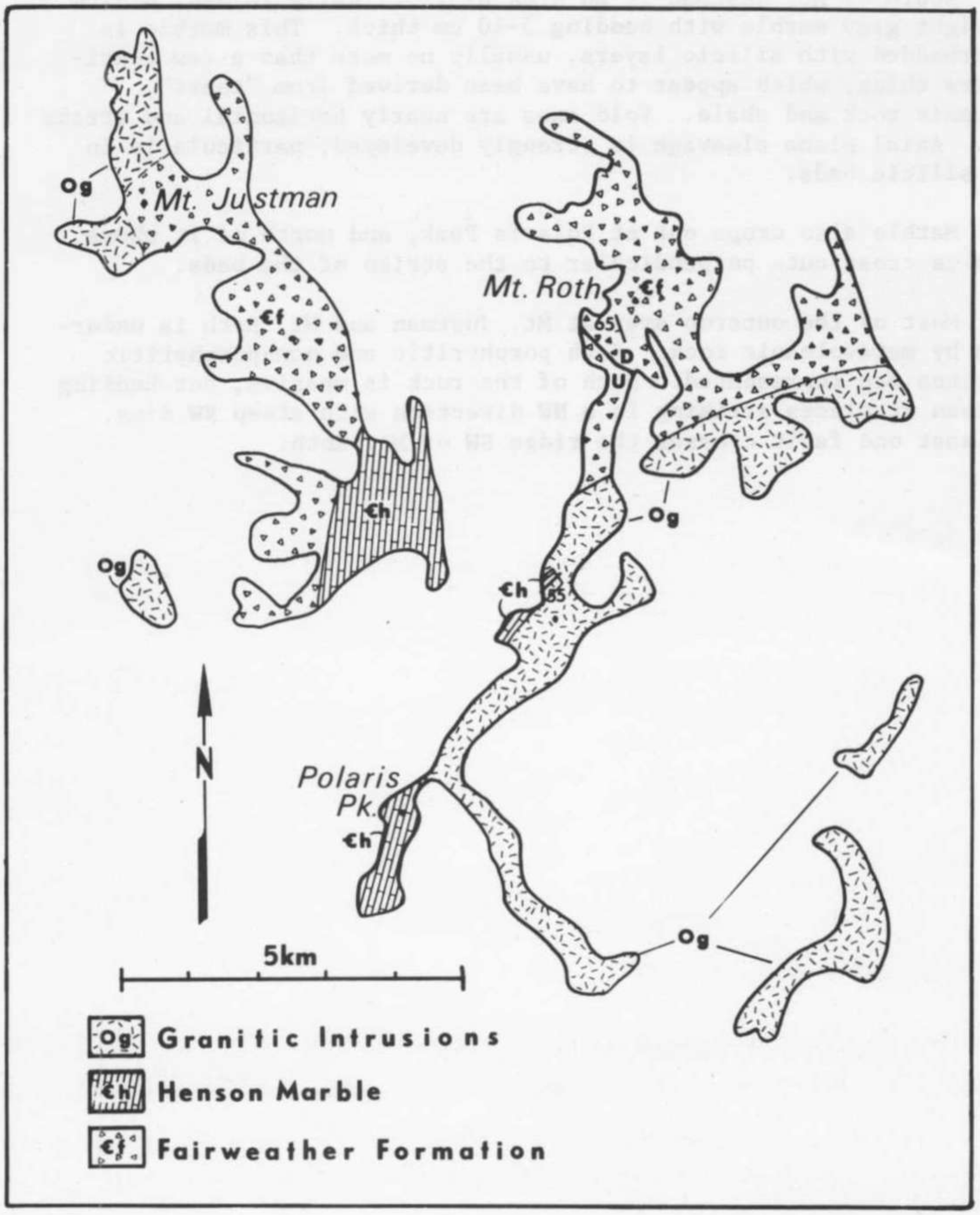
12. Nilsen Peak; 84°32'S, 175°25'W; date: 19 November, 1970.



A zone of mylonitized felsites flanks Nilsen Peak along Massam Glacier. Flaser structures surround the phenocrysts of quartz, plagioclase, and K-feldspar, which in the more deformed rocks are augen shaped. Small amounts of conglomerate or breccia, with clasts 10-20 times longer than wide, were noted. The rocks are bright pink, purple and green, and grade into dark massive porphyries at the west side of Nilsen Peak. This bright coloration is distinctive, with comparable shading found only east of Mt. Fairweather, where porphyritic felsites are sheared, but not mylonized.

The plane of deformation in the mylonites is steeply dipping (75°W to 70°E) about a N-S strike. Directions of movement, as indicated by augen elongation and stretched pebbles, is nearly coincident with the dip of the foliation.

13. Mt. Roth-Mt. Justman; 84°36'S, 172°30'W; dates: 30 November, 1970
16 December, 1970



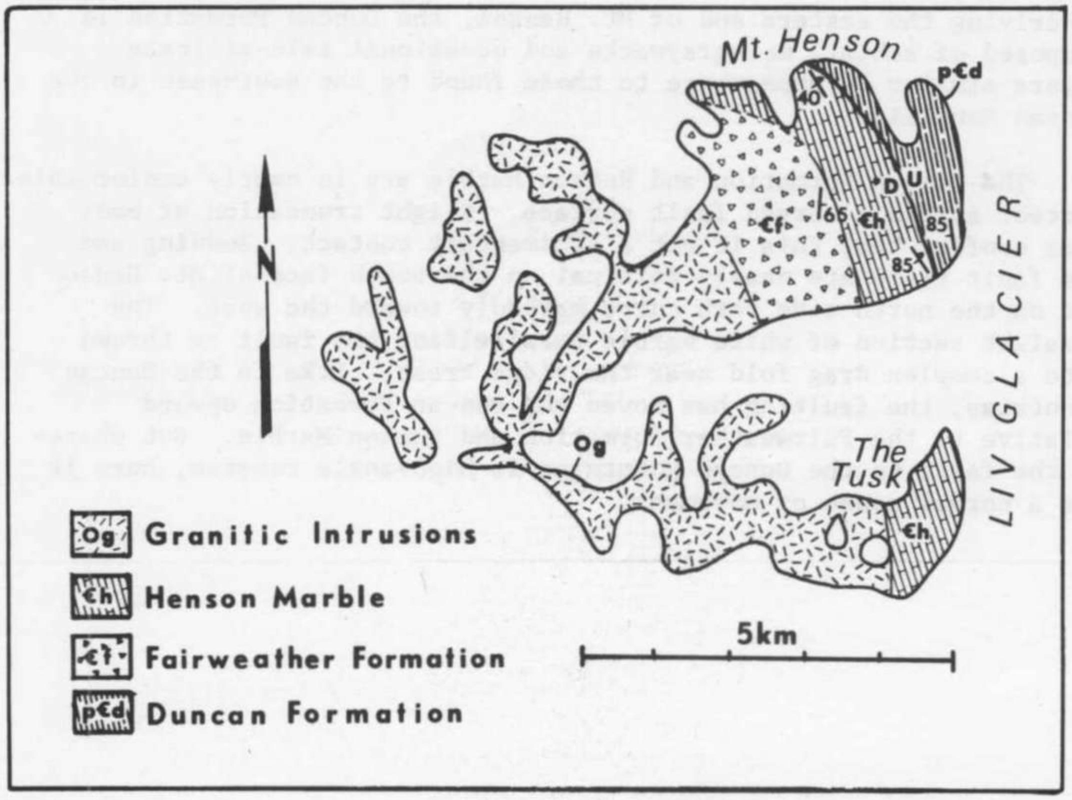
Incomplete reconnaissance has shown this area to be underlain by highly deformed metavolcanic rocks and marbles which are intruded by granite at several places.

South of Mt. Justman is an area of isoclinally folded, medium to light gray marble with bedding 3-10 cm thick. This marble is interbedded with silicic layers, usually no more than a few centimeters thick, which appear to have been derived from "chert", volcanic rock and shale. Fold axes are nearly horizontal and strike NNW. Axial plane cleavage is strongly developed, particularly in the silicic beds.

Marble also crops out at Polaris Peak, and north of it where granite cross cuts perpendicular to the strike of the beds.

Most of the outcrop area at Mt. Justman and Mt. Roth is underlain by metavolcanic rock. Both porphyritic and non-porphyritic felsites are represented. Much of the rock is massive, but bedding is seen at places striking in a NW direction with steep SW dips. At least one fault crosses the ridge SW of Mt. Roth.

14. Mt. Henson; $84^{o}50'S$, $168^{o}18'E$; dates: 12 December, 1970; 29 December, 1974.



Mt. Henson is notable as the informal type locality of the Henson Marble (McGregor, 1965) and for also containing the other two formations of the area, the Duncan and Fairweather Formations. Since the Henson Marble overlies the Fairweather Formation at Mt. Fairweather the same is thought to hold here where the formations are also conformable, although no indicators of this were observed.

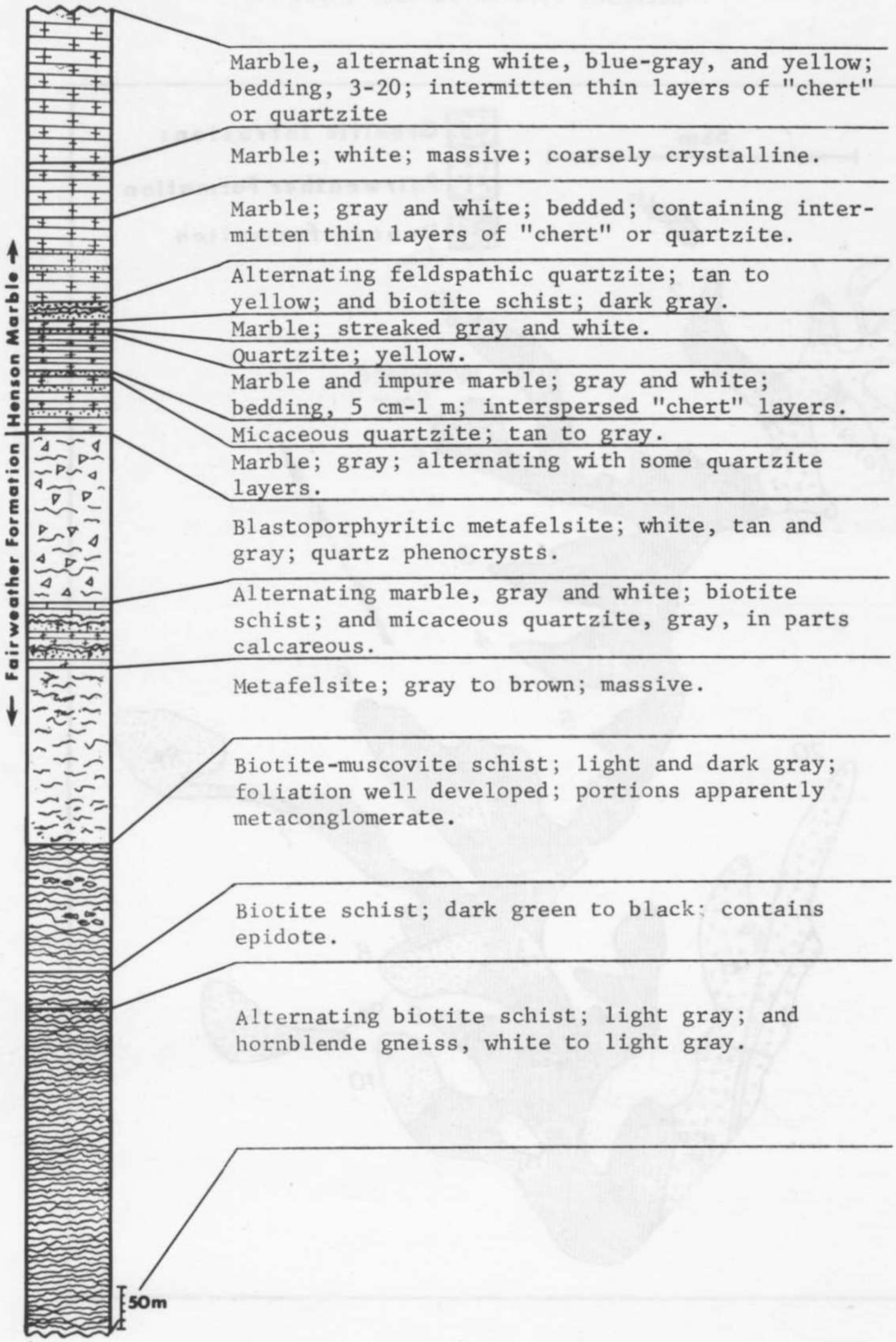
The Henson Marble is a unit of white, coarsely crystalline marble, perhaps 80 m thick, which passes into alternating gray, blue and white beds on either side. It is both highly attenuated and compressed within the exposure area.

Quartzites of the Fairweather Formation interfinger with marble units and then on the ridge to the west of Mt. Henson give way to blastoporphyritic metafelsite, biotite schist and quartzite of which part is calcareous.

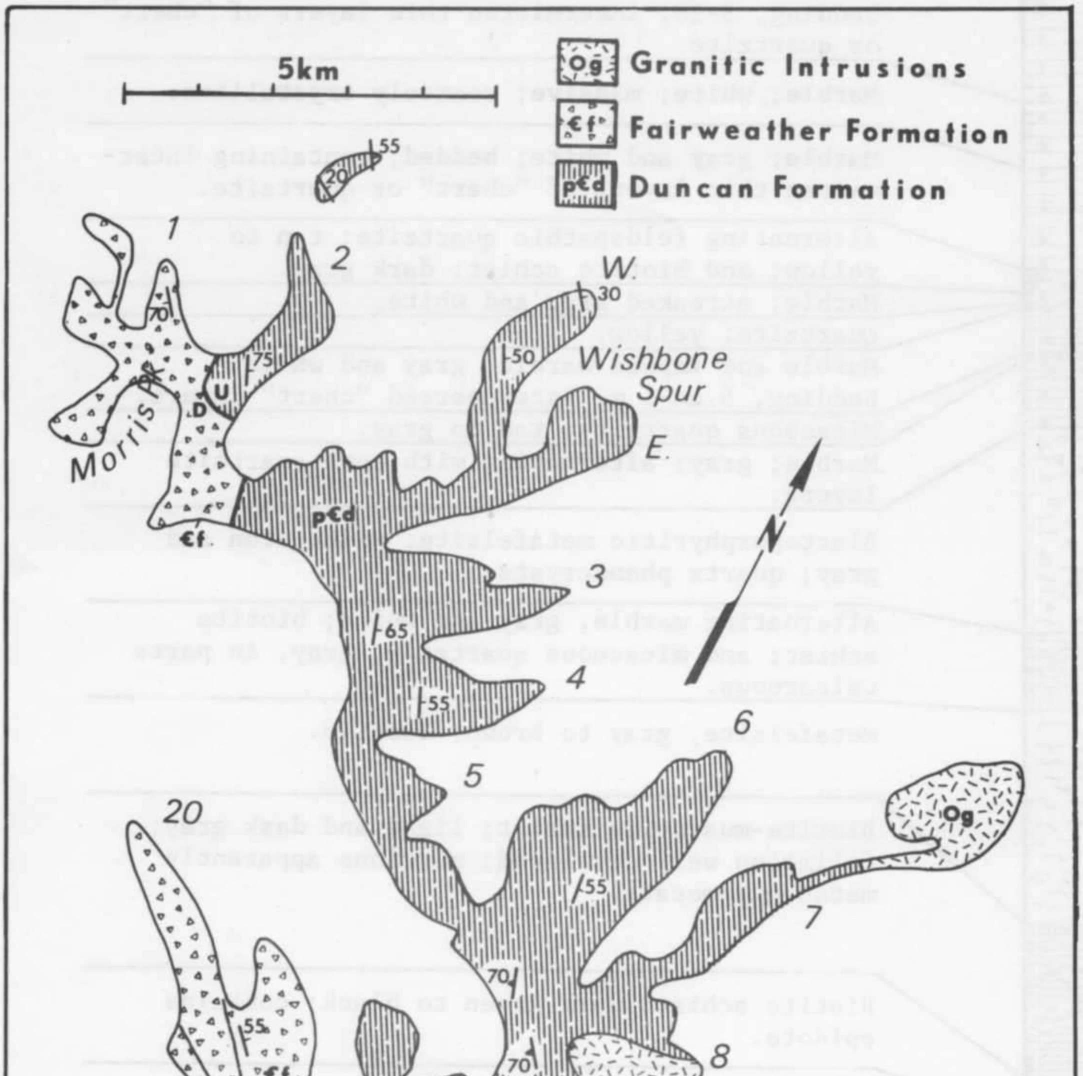
A section was measured along the top of Mt. Henson from the massive white marble westward and at glacier level on the north face from the white marble to the contact with the Duncan Formation. Underlying the eastern end of Mt. Henson, the Duncan Formation is composed of schist, metagraywacke and occasional calc-silicate layers similar in appearance to those found to the southeast in the Duncan Mountains.

The Duncan Formation and Henson Marble are in nearly conformable contact across a curved fault surface. Slight truncation of beds does confirm that this is not a sedimentary contact. Bedding and the fault trace are nearly vertical on the south face of Mt. Henson but on the north side both curve markedly toward the west. The straight section of white marble parallelling the fault is thrown into a complex drag fold near the ridge crest. Like in the Duncan Mountains, the faulting has moved the Duncan Formation upward relative to the Fairweather Formation and Henson Marble. But whereas the fault in the Duncan Mountains is high-angle reverse, here it has a normal sense of movement.

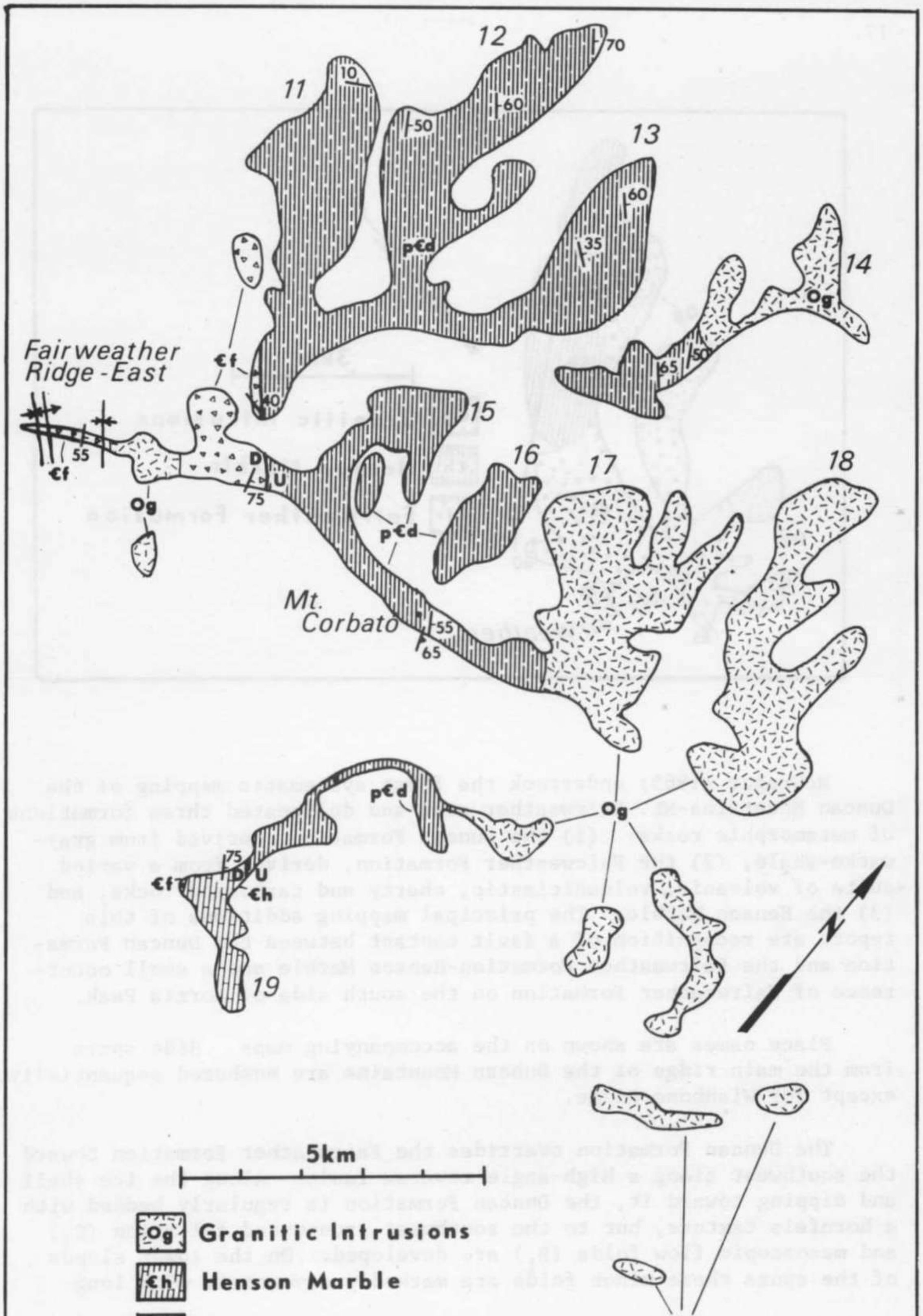
Figure 20. Stratigraphic Section, Henson Marble and Fairweather Formation - Mt. Henson



Figures 15, 16, 17. Duncan Mountains-Mt. Fairweather; 85°00'S, 166°30'W; dates: 13, 14, 15. January 1971, December 1974 to January 1975.



16.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Fairweather Ridge-East
Mt. Corbato
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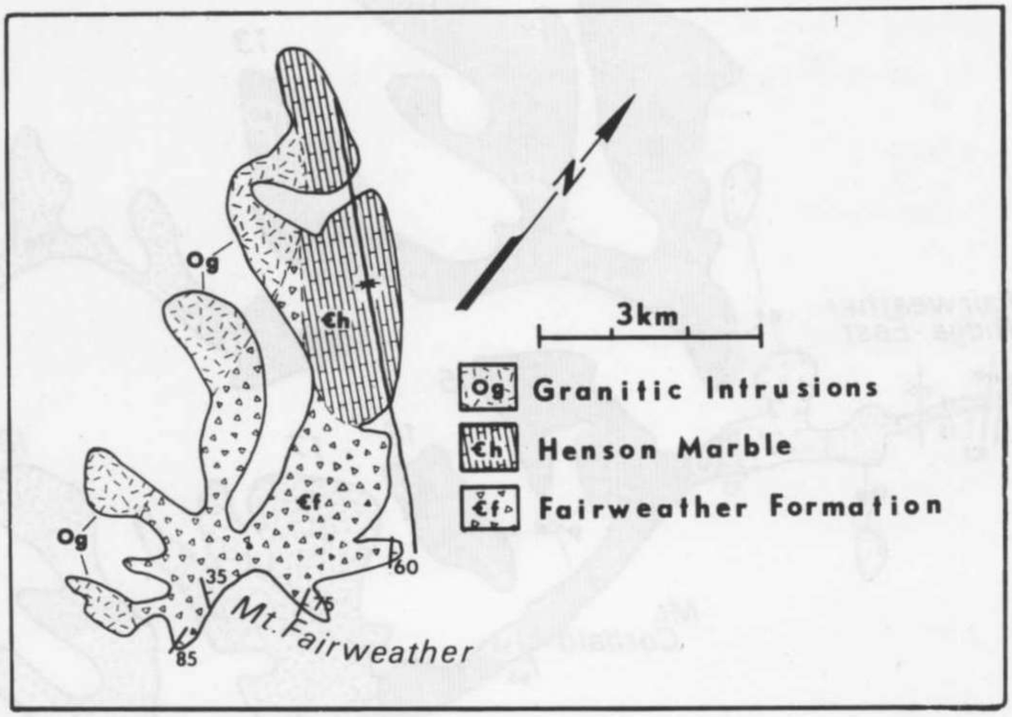
5km
Og Granitic Intrusions
€h Henson Marble
€f Fairweather Formation
p€d Duncan Formation

17.



McGregor (1965) undertook the first systematic mapping of the Duncan Mountains-Mt. Fairweather area and delineated three formations of metamorphic rocks: (1) the Duncan Formation, derived from graywacke-shale, (2) the Fairweather Formation, derived from a varied suite of volcanic, volcaniclastic, cherty and carbonate rocks, and (3) the Henson Marble. The principal mapping additions of this report are recognition of a fault contact between the Duncan Formation and the Fairweather Formation-Henson Marble and a small occurrence of Fairweather Formation on the south side of Morris Peak.

Place names are shown on the accompanying maps. Side spurs from the main ridge of the Duncan Mountains are numbered sequentially, except for Wishbone Ridge.

The Duncan Formation overrides the Fairweather Formation toward the southwest along a high-angle reverse fault. Along the ice shelf and dipping toward it, the Duncan Formation is regularly bedded with a hornfels texture, but to the southwest pronounced foliation (S_1) and mesoscopic flow folds (B_1) are developed. On the lower slopes of the spurs these minor folds are markedly asymmetric with long limbs of anticlines to the northeast, but as the fault is approached on the upper slopes of the main ridge line, the minor folds become symmetric in cross section and axial plane cleavage (S_1) becomes prominent completely obscuring the bedding (S_0) in its extreme development. The axes of these mesoscopic folds (B_1) plunge at about 25 NW. A few kink folds (B_2) were found in the S_1 foliation along the main ridge line northwest of Mt. Corbató.

The Fairweather Formation is tilted northeast toward the Duncan Formation and is asymmetrically folded with sense of overturning toward the southwest. Beds in the Duncan Formation and Fairweather Formation (or Henson Marble) are parallel to one another across the fault contact except on Morris Peak where the Fairweather Formation sits with a marked discordance to the fault. The syncline mapped by McGregor (1965) on the main northwest ridge of Mt. Fairweather contains several anticline-syncline pairs developed in the Henson Marble. It trends across Fairweather Ridge-East, but due to the regional plunge only metafelsites of the Fairweather Formation are exposed there. More work is needed on the northwest ridge of Mt. Fairweather before the structural details there will be understood. An additional asymmetrical pair of folds, overturned to the southwest occurs along the crest of Spur #20. Along Fairweather Ridge-East near the fault contact with the Duncan Formation the Fairweather Formation has been sheared such that pebbles and coarse sand grains are highly attenuated. However this changes northwesterly about 1 km along strike to relatively undeformed outcrops at the base of the ridge. Cleavage is developed only locally in the Fairweather Formation and minor folds were nowhere observed.

The marked contrast in deformational style of the Duncan and Fairweather Formations at all but the largest scale reflects the basic difference in material properties of the constituents of these two formations; namely, pelitic schists of the former and silicic metafelsites and clastics of the latter.

Where the S_1 foliation is not well developed the Duncan Formation can be seen to have very regular beds ranging from about 10 cm to 2 m thickness. Lamination is the most common sedimentary feature and dish structure occurs not uncommonly. A single graded bed with a basal conglomerate (clasts to 3 cm) was observed on the west prong of Wishbone Ridge.

The Duncan Formation has been metamorphosed throughout. Peltic schists and hornfels are the principal rock type with biotite being the predominant mica. Andalusite porphyroblasts to 20 cm occur in the schists northwest of Mt. Corbató Fibrous sillimanite is developed in rocks near some of the granitic intrusions. Colorful brown, green and white layered rocks with calc-silicate assemblages including tremolite/actinolite and sphene are found sporadically, indicating that parts of the Duncan Formation were originally calcareous.

The Fairweather Formation crops out on Mt. Fairweather and Fairweather Ridge-East, as well as on Spur #20, at Morris Peak, and at the tip of Spur #19. McGregor (1965) informally designated the type section along the ridge between Mt. Fairweather and the contact with the Duncan Formation. A composite section was measured in 1975 along the southern and eastern ridge crests of Mt. Fairweather and along Fairweather Ridge-East. Compensation for the snow covered interval occurring between Mt. Fairweather and Fairweather Ridge-East was accomplished by measuring the northeast spur of Mt. Fairweather which covers the gap. The fold in the massive metafelsite at the western end of Fairweather Ridge-East was inferred from the occurrence of mixed chert and metafelsite following 550 m of massive metafelsite on Mt. Fairweather's northeast spur. This is a continuation of the syncline in the Henson Marble on the northwest ridge of Mt. Fairweather. The chert, like the Henson Marble, is not seen on Fairweather Ridge-East due to the regional plunge to the northwest.

The Fairweather Formation exceeds 1340 m. Its lower portion is characterized by basic metavolcanic rocks or greenstone, silicic metafelsite and cherty rock. In the middle portion a thick unit of massive metafelsite is overlain by a sequence of bedded cherty and clastic rocks intercalated with metafelsites. A few cross-beds in this portion indicate top is toward the northeast. An unknown thickness is missing where the middle and upper portions of the section are separated by an intrusion for an 800 m interval underlain by a breccia of dark porphyritic metafelsite floating in white syntectonic granite. The breccia blocks are up to 2 m in length, are exceedingly angular, and are all aligned, with long axes near vertical and parallel to bedding in adjacent strata. Breccia blocks are absent in parts to the east where only the granitic rock crops out. This occurrence is interpreted as the margin of a small pluton intruded into metafelsites of the Fairweather Formation during its deformation.

These rocks are overlain by porphyritic metafelsites and a sequence of clastic and calcareous clastic rocks. These metasedimentary rocks contain trough cross-bedded units, conglomerates and breccias, and wavy, banded beds with greater and lesser amounts of calcareous material.

The occurrences of Fairweather Formation on Spur #20 and at Morris Peak are also clastic and calcareous clastic metasediments. Conglomerates and breccias are common on Spur #20 and an impure marble is infolded on its northeast flank. On Spur #1 the conglomerates are distinctive in that they contain scattered cobbles of orange marble in addition to the quartzitic varieties found elsewhere.

The Henson Marble outcrops at Spur #19 and on the northwest ridge of Mt. Fairweather. It is white and shades of light gray, blue, tan and pink. Interlocking calcite crystals vary from fine- to very coarse-grained. Thin (3-10 cm) beds of silicic rock, possibly of volcanic origin, occur sporadically throughout the marble, often being boudinaged. Bedding in the marble varies from 5 cm to several meters. At Spur #19 the beds are regular or show some undulation from the deformation. Near the fault with the Duncan Formation the marble is extremely sheared with small (less than 1 cm) fragments of silicic material floating in the gray, fine-grained marble. On the northwest ridge of Mt. Fairweather the Henson Marble is highly folded and has a well developed cleavage.

At the tip of Spur #19 several meters of porphyritic metafelsite and some laminated silicic beds occur which probably represent the uppermost part of the Fairweather Formation at this location.

Granitic plutons intrude an appreciable portion of the Duncan Mountains as indicated on the accompanying maps.

Figure 21. Stratigraphic Section, Fairweather Formation- Mt. Fairweather and Fairweather Ridge-East.

The section is measured from the top in five parts as numbered in Figure 4.

Section No. 1

35 m	Quartzite, gray
12	Quartzite, gray and tan, green lenses.
40	Gritty quartzite.
1	Conglomerate.
10	Mixed conglomerate and quartzite.
<1	Quartzite, green, with amphibole needles.
40	Quartzite, gray, lenses of conglomerate.
6	Quartzite, gray, with calc-silicate areas.
15	Quartzite, gray, with calcareous layers.
16	Quartzite, with coarse lamination, occasional calc-silicate area.
1	Conglomerate.
35	Gritty quartzite, with sparse pebbles.
5	Metafelsite, gray to black, some calc-silicate areas.
3	"Chert", brown streaks.
2	Metafelsite, gray to black.
<1	Gray marble.
5	Metafelsite, gray to black.
<1	Marble, with laminations.
26	Metafelsite, gray to black.
35	Quartzite, gray, tan, green.
<1	Sheared marble.
6	Quartzite, tan, sheared.
6	Blastoporphyritic metafelsite, orange.
1	"Chert", red.
6	"Chert", gray, sheared.
62	Blastoporphyritic metafelsite, orange, tan, yellow.
9	Metafelsite, pink and green.
52	Blastoporphyritic metafelsite.
3	"Chert", gray with white spots.
128	Blastoporphyritic metafelsite, gray.
18	Blastoporphyritic metafelsite, pink.
123	Breccia of gray, blastoporphyritic metafelsite in granite.
180	Granite with occasional zones of breccia.
450	Breccia of gray, blastoporphyritic metafelsite in granite.
5	Breccia with small fragments.
10	Breccia of gray, blastoporphyritic metafelsite in granite.
28	Blastoporphyritic metafelsite.
16	Breccia of blastoporphyritic metafelsite in granite.

Section No. 1

8 m	"Chert", green, banded.
3	-Covered-
13	"Chert", green, banded.
7	"Chert", green and gray.
5	Blastoporphyritic metafelsite, with garnets.
<1	Garnet schist.
1	Blastoporphyritic metafelsite.
1	Garnet schist.
3	Amphibolite.
2	"Chert", speckled.

Section No. 2

38 m	"Chert", dark gray.
4	"Chert", greenish gray.
2	-Covered-
37	"Chert", dark gray-green.
5	Amphibolite.
54	"Chert", gray-green.
3	-Covered-
20	Metafelsite, black, foliated.
7	"Chert", dark gray to black.
16	Blastoporphyritic metafelsite, dark gray to black.
2	"Chert", dark gray.
4	Amphibolite.
2	"Chert", speckled.

Figure 4 (continued)

Section No. 3.

12 m	"Chert", speckled.
10	"Chert", dark gray.
4	Sheared metafelsite.
32	-Covered-
46	"Chert", dark gray.
8	Metafelsite, dark gray, bedded.
1	"Chert", tan.
3	Metafelsite.
3	"Chert" and marble.
4	"Chert", tan and gray.
35	Metafelsite, with small crystal fragments.
15	"Chert", black.
21	Metafelsite, green and black streaks.
12	Blastoporphyritic metafelsite, foliated.
3	Amphibolite.
21	Blastoporphyritic metafelsite, black.
21	Blastoporphyritic metafelsite, red to yellow.
4	"Chert", green to black.
25	Amphibolite, green to black.
19	Metafelsite, green and black streaks.
9	Blastoporphyritic metafelsite, red-gray.
32	Blastoporphyritic metafelsite, greenish gray.
20	Blastoporphyritic metafelsite, red-gray.
11	"Chert", red-gray, bedded.
32	"Chert", reddish black.
4	Biotite-hornblende schist.
<1	Amphibolite.
3	Blastoporphyritic metafelsite, red.
7	Volcanic breccia, gray.
23	"Chert", black to gray.
<1	"Chert", yellow.
4	"Chert", black.
14	Foliated greenstone.
19	"Chert", black and yellow.
36	"Chert", laminated.
336	Blastoporphyritic metafelsite, red, black, gray.
4	Blastoporphyritic metafelsite, black.
39	Blastoporphyritic metafelsite, gray to tan.
18	Blastoporphyritic metafelsite, black and green.
54	Blastoporphyritic metafelsite, gray.
101	Blastoporphyritic metafelsite, gray to brown.

Figure 4 (continued)

Section No. 4.

	"Chert", massive and bedded.
510 m	Blastoporphyritic metafelsite, gray, orange, tan.
40	Blastoporphyritic metafelsite, dark gray to black.
	Volcanic Breccia.

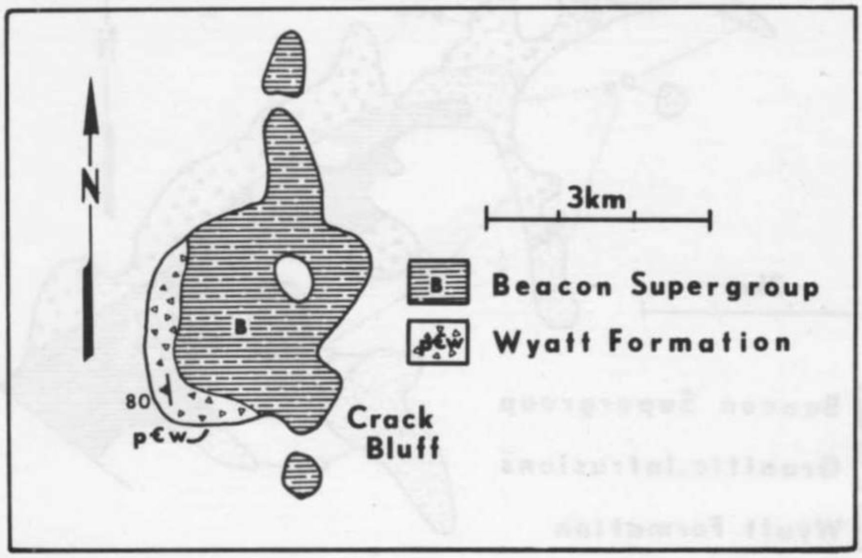
Section No. 5.

114 m	Blastoporphyritic metafelsite, gray.
10	-Covered-
25	Blastoporphyritic metafelsite, gray.
7	-Covered-
9	Blastoporphyritic metafelsite, gray.
45	Blastoporphyritic metafelsite, green and red.
200	Volcanic breccia.
5	Greenstone.
25	Volcanic breccia.
74	Greenstone.
33	"Chert", red, green, gray.
18	Greenstone.
40	Amygdaloidal metabasalt.
28	"Chert", spotted.
38	Metafelsite, with fine crystal fragments.
6	Greenstone.
4	"Chert", red, black, light green.
10	Blastoporphyritic metafelsite, black.
4	"Chert", light green streaks.
15	"Chert", spotted.
93	Blastoporphyritic metafelsite, black.
52	Metafelsite, black.
18	Amphibolite.
27	"Chert", black, brown, tan.
6	Amphibolite.
4	Marble.
12	Basaltic breccia.
3	Blastoporphyritic metafelsite.
10	Amphibolite.
18	Basaltic breccia.
65	Blastoporphyritic metafelsite, brown.

Western Nilsen Plateau; $86^{o}00'S$, $159^{o}00'W$.

Exposures on the western side of Nilsen Plateau include a sequence of metamorphosed graywacke and shale and a massive porphyritic felsite, probably correlative, respectively, with the LaGorce and Wyatt Formations of the southern Scott Glacier area 60 km to the east. Most of the southwestern flank of the plateau is underlain by granodiorite and quartz monzonite, with discrete plutons intruding the metamorphic rocks on the northwestern flank. North of Moraine Canyon migmatites are present intimately associated with granitic rocks. The steep upper slopes of Nilsen Plateau are of late Paleozoic Beacon strata unconformably overlying the basement complex.

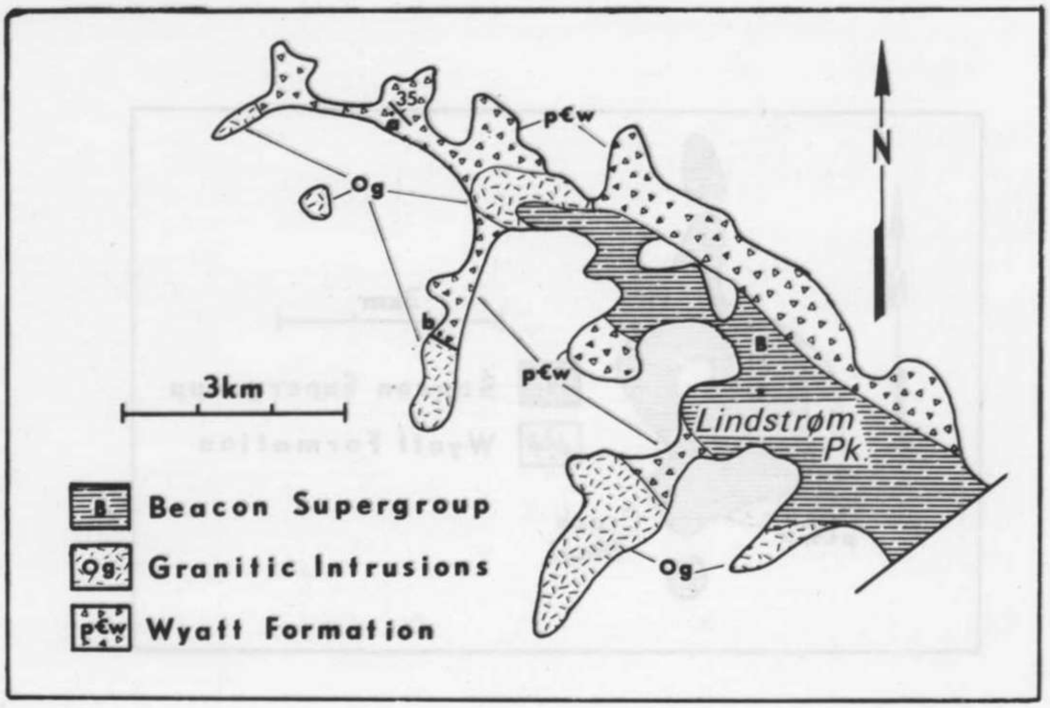
18. Crack Bluff; $86^{\circ}33'$S, $158^{\circ}50'$; date: 27 December, 1970.



An isolated outcrop of porphyritic felsite occurs beneath the Beacon strata at the southermost tip of Nilsen Plateau. The rock is massive and dark gray with phenocrysts of embayed quartz (to 4 mm) and highly altered plagioclase (to 2 mm) set in a sericitized groundmass. A foliation is developed in much of the rock at about $10^{\circ}85^{\circ}$W.

A subhorizontal breccia sheet several meters thick and several tens of meters long is present. The sharply angular breccia blocks and fragments are of identical composition to the enclosing felsite and indicate a brittle mode of fracture. These fragments are enclosed in a quartz druse which in places failed to fill the wider interstricies. Because of this, the brecciation surely post-dates the Ross Orogeny, and it seems likely that it also would have followed burial by the Beacon sediments. Steeply upturned Beacon beds indicate that the southwestern edge of Nilsen Plateau traces the locus of a monocline upthrown and eroded away on the southwest, thus warranting the interpretation that this subhorizontal breccia body is a brittle, tensional separation which occurred during development of the fold.

19. Lindstrøm Peak; $86^{\circ}18'S$, $160^{\circ}10'W$; date: 29 December, 1970.

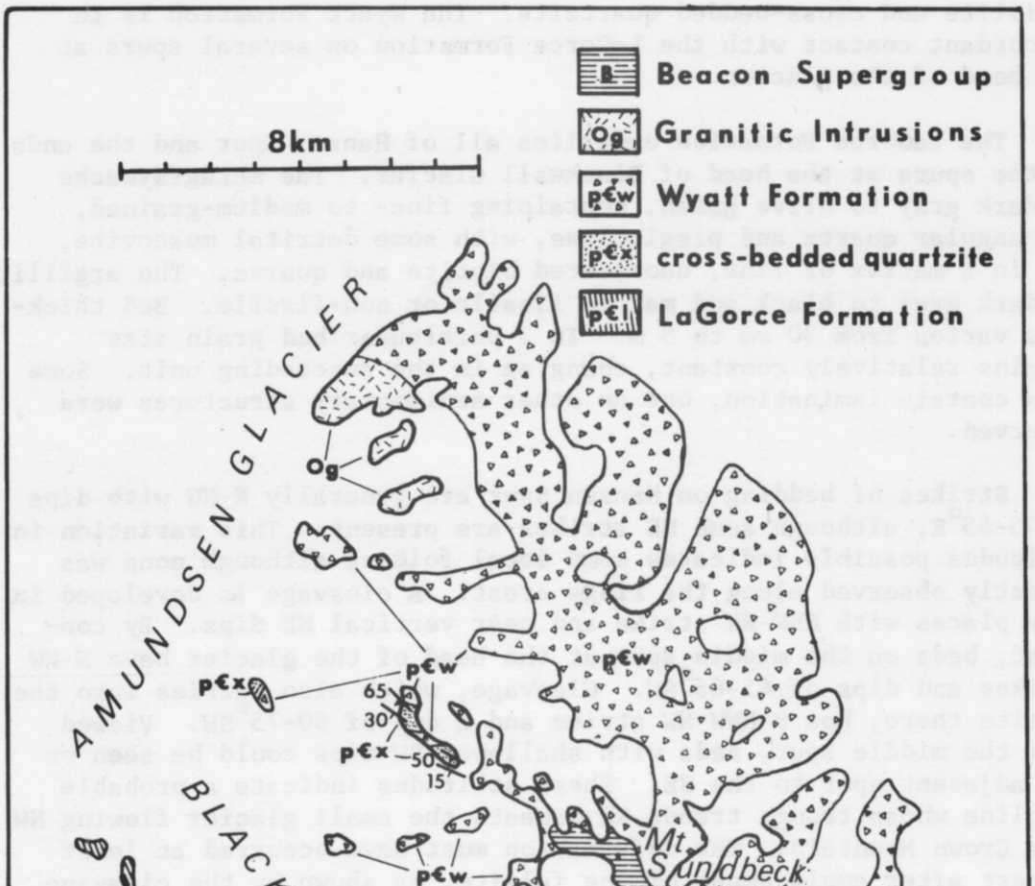


The spurs to the northwest of Lindstrøm Peak are underlain by massive porphyritic felsites and granitic rock. Colors of the felsite are variable, including gray, pink and green. The numbers and type of phenocrysts also varies, with quartz-plagioclase-biotite at some places and quartz-plagioclase-K-feldspar at others. The plagioclase is highly altered to epidote and biotite remains as alternating layers of chlorite and opaque minerals.

Near point a the felsite exhibits a layering in which phenocrysts are more and less concentrated. There are also thin (1 cm thick) lenses of chert-like material, perhaps felsite without phenocrysts, which indicate bedding strikes at 135° and dips 35°NE.

Veins of epidote and calcite cut through portions of the rock on the southwestern spurs. At least four shear zones of crushed rock are present crossing the spur at point b. McLelland (unpub.) interpreted the contact between the quartz monzonite and the felsite as a fault. No evidence has been found to contradict this interpretation.

20. Blackwall Glacier Area; 86°12'S, 159°00'E; dates: 28 December, 1970; 2, 3, 5, 8, January, 1971.



The Blackwall Glacier area is important in that three distinct rock groups are present: evenly bedded metagraywacke-argillite (LaGorce Formation), porphyritic felsite (Wyatt Formation) and argillite and cross-bedded quartzite. The Wyatt Formation is in discordant contact with the LaGorce Formation on several spurs at the head of the glacier.

The LaGorce Formation underlies all of Hansen Spur and the ends of the spurs at the head of Blackwall Glacier. The metagraywacke is dark gray to olive green, containing fine- to medium-grained, sub-angular quartz and plagioclase, with some detrital muscovite, set in a matrix of fine, unoriented biotite and quartz. The argillite is dark gray to black and may be fissile or non-fissile. Bed thickness varies from 30 cm to 5 m. In a particular bed grain size remains relatively constant, changing in the succeeding unit. Some beds contain lamination, but no other sedimentary structures were observed.

Strikes of bedding on Hansen Spur are generally N-NW with dips of 45-65°E, although some NE strikes are present. This variation in attitudes possibly indicates some local folding although none was directly observed along the ridge crest. A cleavage is developed in many places with NNW-NW strike and near vertical NE dips. By contrast, beds on the middle spur at the head of the glacier have N-NW strikes and dips of 45-65°SW. Cleavage, which also carries into the felsite there, has a NNW-NW strike and a dip of 60-75°SW. Viewed from the middle spur, beds with shallower SW dips could be seen on the adjacent spur to the SE. These attitudes indicate a probable syncline whose trough trends SE beneath the small glacier flowing NW from Crown Mountain. The deformation must have occurred at least in part after emplacement of the felsite, as shown by the cleavage.

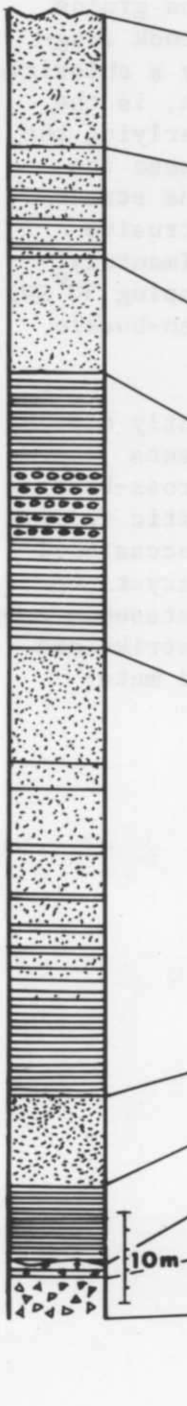
The near vertical contact between the felsite and metasediments is more likely intrusive than unconformable, since an elaborate structural sequence including overturning would have to be invoked to produce the present relationship. Whereas, the LaGorce rocks appear to have been folded only once. The contact is knife-sharp. No alteration zone is apparent in the metasediments, although in the felsite near the contact there is an absence of phenocrysts, and the presence of some lithic fragments.

On the spurs at the head of Blackwall Glacier the dark gray to brown felsite has an extremely fine microcrystalline groundmass of quartz, feldspar(?), and sericite, biotite, and chlorite. Phenocrysts of biotite, euhedral twinned plagioclase, and rounded and embayed quartz vary in abundance throughout the rock. In one specimen concentrations of crystal fragments exist between areas of groundmass and unbroken phenocrysts. In other cases the majority of phenocrysts are unfractured. Felsite interspersed with zones of cleavage underlies most of the eastern side of the valley, as well as much of Moraine Canyon to the northeast.

On a ridge toward the middle of the area there exists a 150 m section of cross-bedded quartzite and argillite with some globular "chert" layers (see Figure 22). The quartzites contain angular to sub-angular quartz and some plagioclase. Pyrite and zircon grains define laminations in the festoon cross-beds. Phyllitic rock fragments were discernable in some samples, but in others only a chloritic matrix was apparent. A highly sheared marble, 1-2 m thick, is the lowest bed and is in contact with porphyritic felsite underlying the rest of the ridge. Movement appears to have occurred between the sediments and felsite with the marble absorbing most of the strain. But whether the conformable contact is depositional or intrusive parallel to bedding could not be determined. The metasedimentary rocks dip gently to the north and cleavage (NW strike, dipping 50°SW) is developed in some of the finer-grained beds. Small open-buckle folds occur in some layers.

On the end of the ridge northwest of Mt. Sundbeck gently dipping argillite and graywacke crops out, more akin to the sediments just described than to the LaGorce Formation, though without cross-bedding. These sediments are intruded by a dark gray-green porphyritic felsite distinguished by quartz phenocrysts with a pink cast and occasional glomeroporphyritic aggregates of nearly pure quartz phenocrysts up to 15 cm in diameter. Several shear zones occur in the metasediments and along the contact at places. Cleavage with NW strike and steep SW dip carries through parts of both the felsite and metasediments.

Figure 22. Stratigraphic Section, Cross-bedded quartzite - Black Rock Glacier



Sandstone, fine to medium-grained; massive; cross-beds to 0.5 m; clay galls in few layers.

Shale and fine sandstone; black to dark gray; bedding, 5-15 cm; alternating with green "chert" layers; bedding, 3-8 cm.

Quartzite; tan to light gray; fine to medium-grained; bedding, 2-3 m; faint cross-bedding.

Upper portion: Shale to fine sandstone; bedding, 5-20 cm; lamination in upper part.

Middle portion: Shale to fine sandstone; black; "chert" blobs with 10-20 cm spacing.

Lower portion: Shale; black; some mica flakes; bedding about 1 m.

Sandstone; fine to medium grained; laminated; few cross-beds;

changes downward into

Shale; black; bedding, 0.5-1 m; with "chert" layers; 2-4 cm thick; continuous and discontinuous.

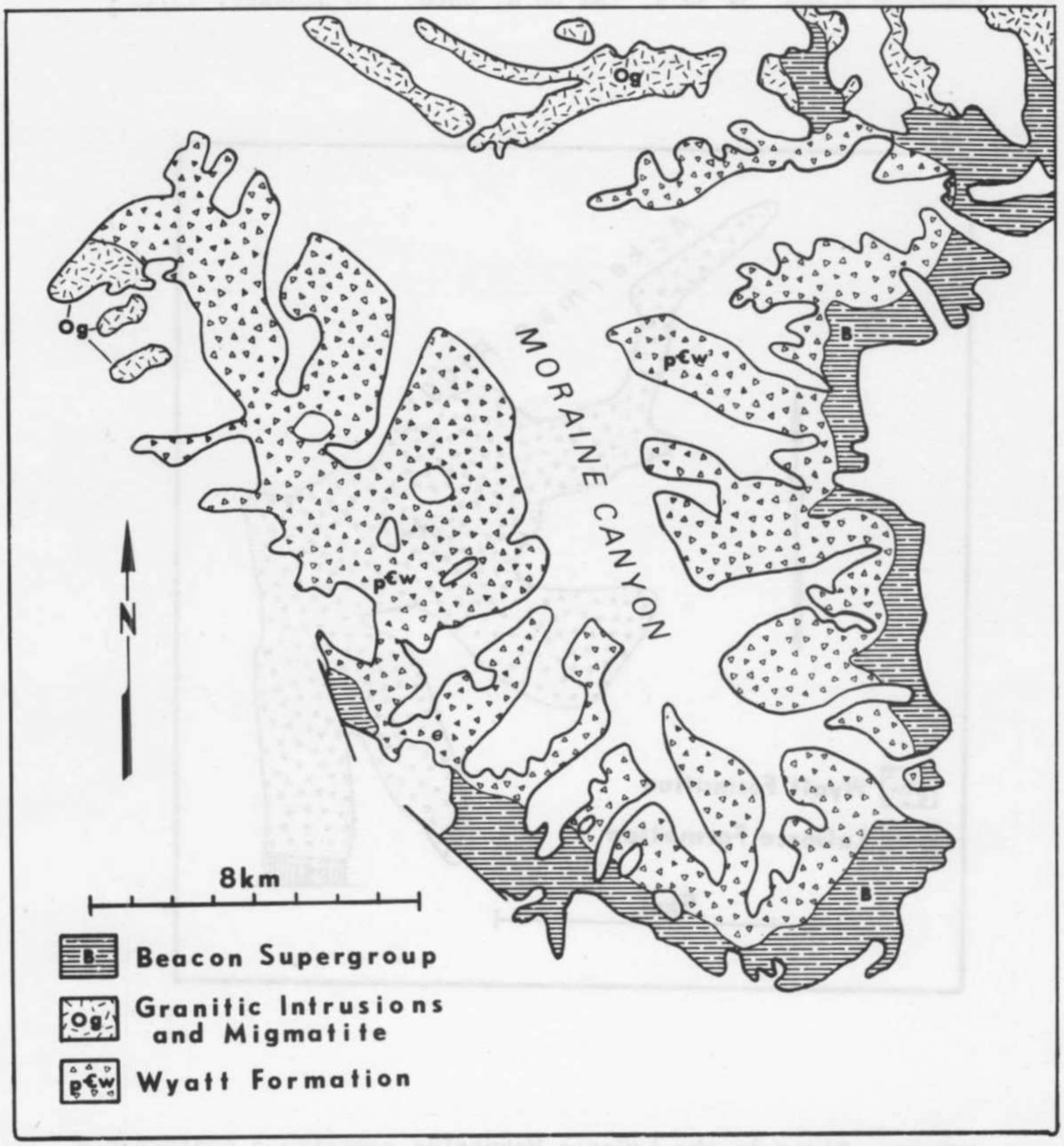
Quartzite; light gray; coarse-grained; bedding, about 2m; cross-bedding, sets, 4-8 cm.

Shale; black; bedding, 0.5-10 cm; laminated in bottom 1/3.

Marble; highly sheared; white; medium-grained; containing stringers of dark "chert".

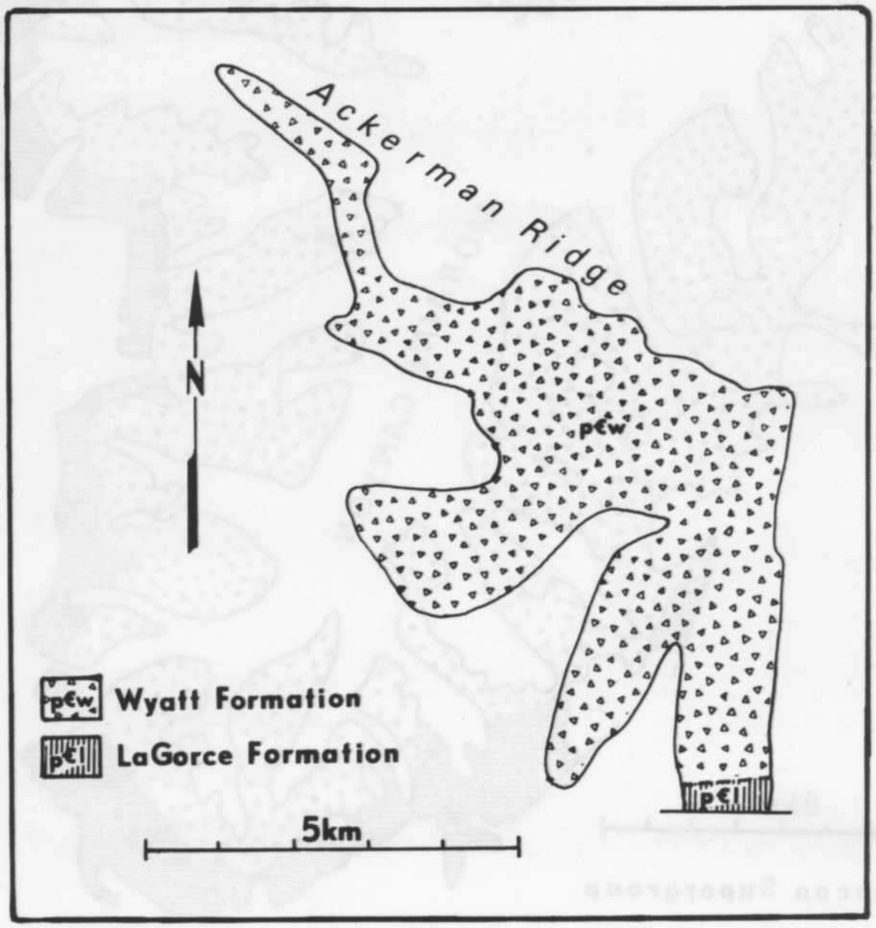
Porphyritic felsite; dark gray; quartz phenocrysts.

21. Moraine Canyon; 86°10'S, 158°00'W; dates: 31 December 1970; 19 January, 1971.



Moraine Canyon is underlain by a continuation of the dark, porphyritic felsites found to the south in the Blackwall Glacier area. The rocks are massive and appear similar at all outcrops visited. Petrographically they are similar to the felsite of the Blackwall Glacier area with embayed quartz and altered plagioclase and biotite crystals and fragments set in a microcrystalline groundmass of quartz, feldspar and mica. In the groundmass plagioclase concentrations occur at places, K-feldspar at others. Most of the samples also contain areas of chlorite, thought to be altered basic igneous rock fragments.

22. Ackerman Ridge; 86°34'S, 184°00'W; date: 18 January, 1971.



Ackerman Ridge in the LaGorce Mountains containa a conformable contact between the LaGorce and Wyatt Formations (Katz and Waterhouse, 1970a). The LaGorce Formation is a sequence of alternating metagraywacke and argillite. Beds are 0.5-3 m in thickness and color is mostly gray-green with some beds gray and tan.

The LaGorce Formation is overlain by a massive; greenish, porphyritic felsite of the Wyatt Formation, perhaps 500 m thick. This unit is highly foliated for about 200 m adjacent to the contact which is near vertical and strikes E-W across the ridge. The thick felsite is overlain by a mixed sequence of volcanic, volcaniclastic, and clastic rocks. The clastics include purple and green argillite, graywacke and fine- to very coarse-grained quartzites, some of which are cross-bedded, and indicate tops to the north, confirming Minshew's (1967) suggestion that the Wyatt Formation is younger than the LaGorce.

There are a few layers of sparse-cobble conglomerate with occasional rounded quartz cobbles (5-15 cm) set in a matrix of coarse sand. Felsite units are usually only 5-10 m thick although some are considerably thicker. This mixed sequence is overlain by more massive porphyritic felsite that extends northward on the ridge.

Structurally the LaGorce Formation is complexly folded at a distance from the contact. The Wyatt Formation exhibits small folds in some layers but on a gross scale does not appear to be folded. Strikes are E-SE with near vertical to steeply overturned dips.

APPENDIX B. HISTORICAL BACKGROUND--PRE-CAPE ROCKS,

CAPE PROVINCE, SOUTH AFRICA

Introduction

The Geological Commission of the Cape of Good Hope commenced work on the western and central portions of the province in 1895. The majority of the effort for the first three years, and to a lesser degree after that, was on surveying the pre-Cape rocks, a group unconformably overlain by the Mid-Paleozoic Cape System. A. W. Rogers, E. H. L. Schwarz, and G. A. Corstorphine published numerous progress reports. By 1911 this information had been compiled on four geological map sheets covering the western Cape (Rogers and others, 1906a; 1906b; 1911; Rogers and Schwarz, 1907). The name "Malmesbury" was retained from the usage of E. J. Dunn (1873) on his sketch map of Cape Colony for the folded and intruded slates and grits of the southwestern districts (Rogers, 1896). The rocks of the Malmesbury were found generally to be clay-slate, phyllite, mica schists, and quartzite, often with a well developed cleavage. Several NW-SE trending groups of granite plutons were also delineated (Corstophine, 1897).

Western Cape--Early

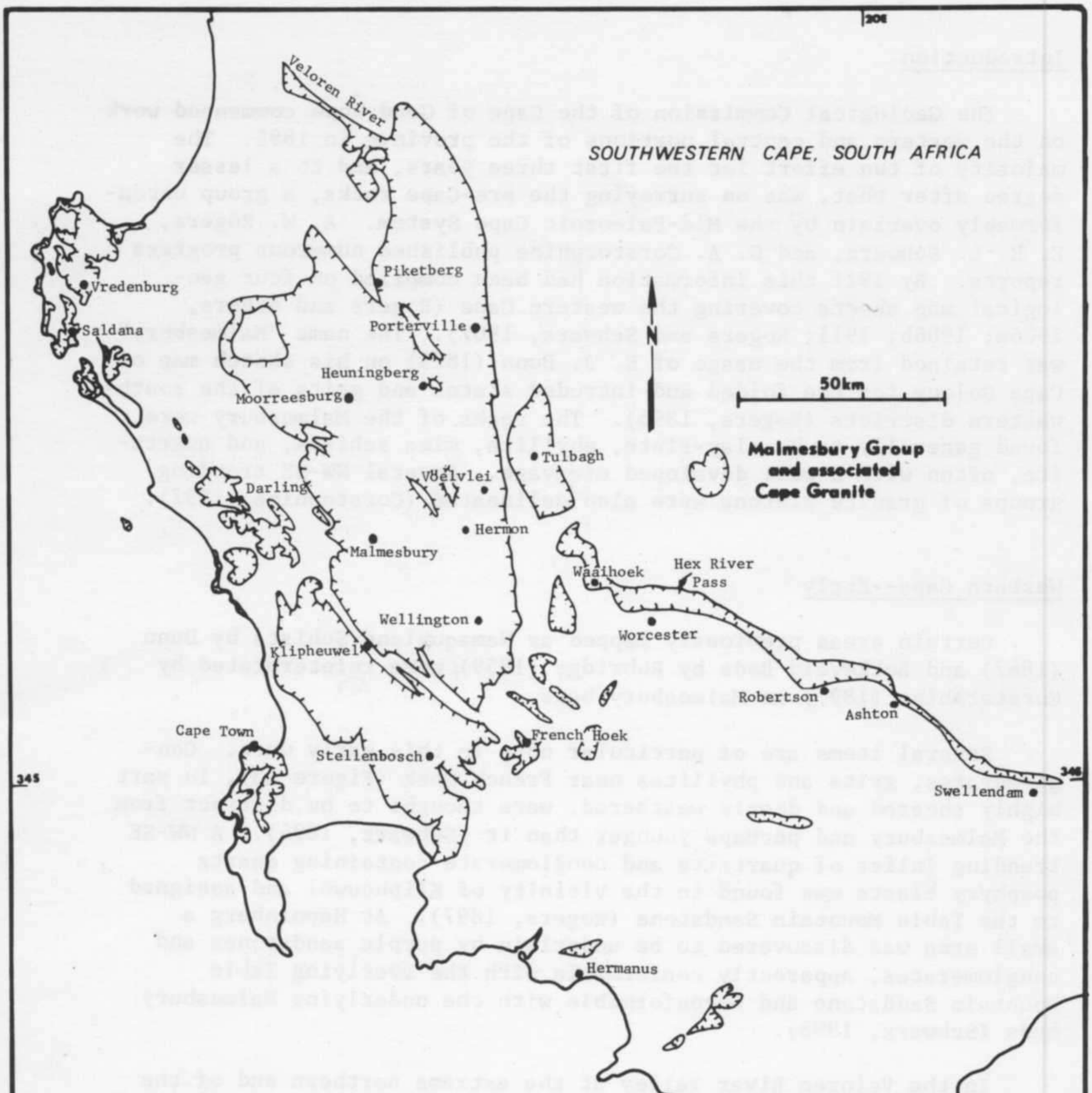
Certain areas previously mapped as Namaqualand Schists by Dunn (1887) and Bokkeveld Beds by Rubridge (1859) were reinterpreted by Corstorphine (1897) as Malmesbury beds.

Several items are of particular note in this early work. Conglomerates, grits and phyllites near French Hoek (Figure 23), in part highly sheared and deeply weathered, were thought to be distinct from the Malmesbury and perhaps younger than it (Schwarz, 1896). A NW-SE trending inlier of quartzite and conglomerate containing quartz porphyry clasts was found in the vicinity of Klipheuwel and assigned to the Table Mountain Sandstone (Rogers, 1897). At Heuninberg a small area was discovered to be underlain by purple sandstones and conglomerates, apparently conformable with the overlying Table Mountain Sandstone and unconformable with the underlying Malmesbury beds (Schwarz, 1898).

In the Veloren River Valley at the extreme northern end of the Malmesbury outcrop area are two small exposures of less deformed, red and green shale and sandstone with detrital mica (Rogers, 1903). These were correlated with the Ibiquas Series, described in the Van Rhyn's Drop area to the north (Rogers and Schwarz, 1900) (Figure 14).

Work on the window of Malmesbury beds exposed in the strip from Worchester to Swellendam showed that the rocks are predominantly

Figure 23. Locations, Southwestern Cape, South Africa.



sheared slates with lesser sandy beds, but limestone and calcareous black shale were also found, particularly toward the east (Schwarz, 1896; 1897; 1905b; Rogers and Schwarz, 1898c). These sediments were intruded at two places by plugs of granite.

At this same time work was being carried out on pre-Cape rocks farther to the east in the Cango Valley and in the George area.

Cango--Early

The first foray into the slates, grits, sheared conglomerates and limestones of the Cango Valley was to select a deep borehole site for artesian water (Corstorphine, 1896a). The Cango Cave, formed in a thick dolomitic limestone within the clastic sequence, was also described (Corstorphine, 1896b). In 1898, Rogers and Schwarz completed their survey of the Cango describing the succession throughout the valley. They refer to the rocks collectively as the Malmesbury Beds and distinguish the Cango Conglomerate, composed of rounded pebbles of vein quartz and lesser quartzite, slate, phyllite, and gneiss in a slatey grit matrix. Slates are the most common rock type but intercalations of limestone and quartz-feldspar grit or porphyry are prevalent particularly in the north-central portion of the area. A peculiar layer of conglomerate containing boulders of limestone, granite, and mica schist was also noted (p. 72-73). Citing this in his summary report, Corstorphine (1898), with proper reservations, suggested that the granite clasts indicate the Cango may be younger than the Malmesbury.

George--Early

The initial investigation of the pre-Cape window in the George district found coarse-grained, muscovite-bearing granite predominating in the area, with Malmesbury beds surrounding it (Rogers and Schwarz, 1898a). The section of Victoria Bay showed feldspathic quartzites, sandstones, quartz-phyllites, and massive and banded slates, some being graphitic. Elsewhere the rocks were seen to be the more normal clay-slates and phyllites of the Malmesbury.

After further study of these pre-Cape rocks, Schwarz (1905a) hesitated to correlate them with the Malmesbury of the Western Cape due to their distinctive lithology. He later described the petrology of the andalusite found throughout many of these rocks (Schwarz, 1907).

Rastall and Shand

Following publication of the quadrangle maps by the Geological Commission (1906-1911) little work was done on the pre-Cape rocks until the 1930's when detailed studies were undertaken in most of the areas. Two notable exceptions are the investigations of Rastall (1911) and Shand (1917).

Rastall, on an extended vacation from England, undertook a study of the geology between Worcester and Ashton. His principle contribution was to distinguish in the section north of Worcester and at the Hex River pass, a lower gritty series exposed to the north and an upper slaty series exposed to the south. The grits are composed of quartz, detrital mica, and in some cases feldspar. Accessory heavy minerals are also present. He reports a gradual transition to the finer-grained slaty series in which feldspar appears to be more abundant. The bands of marble north of Worcester were also noted.

Shand (1917) discussing the local geology around Stellenbosh, noted, in addition to the Malmesbury Series, a sequence of sheared grits and conglomerate which he assigned to the French Hoek Series.

Textbooks

Numerous South African geologists of this century have compiled textbooks on the geology of the Republic. Not only are these a reflection of the correlation trends of the time, but often they served as a forum for arguing the authors' opinions on details of Malmesbury stratigraphy and field relations.

Two syntheses of South African geology appeared in 1905, one coauthored by the former Director of the Geological Survey and the other by the Director at the time.

Hatch and Corstorphine (1905) stated that "the oldest rocks--the Malmesbury Series of the south and the Namaqualand and Swaziland Series of the north and east are on the same geological horizon," all of which were then assigned to the Archean System. Mention was made of equivalents of the younger Cango Series occurring in the Gamtoos Valley. Rogers (1905) discussed field occurrences throughout the western and central Cape, choosing to place the rocks of the Gamtoos area with the Malmesbury of the western Cape and George. Rogers and du Toit (1909) in the second edition introduced the name French Hoek for the beds described by Schwarz (1898b) around that town and correlated then with the Cango beds and Ibiquas (younger than the Malmesbury) along with series in more far reaching areas of southern Africa, differing only in a few points from Hatch and Corstorphine (1905) (Appendix C).

On the Union map sheet of 1925 the Malmesbury and Cango again found themselves with the same color, different than the French Hoek, which included Klipheuwel, and altogether under the heading of Transvaal or Nama System (Rogers, 1925).

The following year du Toit's (1926) first edition of The Geology of South Africa appeared. He suggested (p. 49) that the George granite may have originated at the time of the other Primitive Systems. But he also discussed the schists in terms of the Malmesbury (p. 123), included within the Nama System. More significant, however, is that he distinguished the beds unconformable on Malmesbury in the strip northwest of Klipheuwel Station from the beds at French Hoek, and compared them with the beds found at Heuingberg, both of which were speculated to belong to the Masap System. Du Toit also reconciled the two positions held for the Cango by distinguishing the Cango Beds from the Cango Conglomerate and suggesting correlation of the former with the Malmesbury and the latter with the French Hoek.

Schwarz's (1928) textbook grouped all of the pre-Cape rocks of the southern Cape as Malmesbury, belonging to the Nama and Transvaal Systems. Hall (in Rogers and others, 1929) followed this scheme but distinguished the French Hoek beds.

1930's

During the 1930's mapping projects emphasizing pre-Cape rocks were undertaken in the western Cape, Cango Valley and Gamtoos area. Haughton (1932; 1933) followed du Toit's 1926 scheme in his mapping of the new Cape Town sheet. He also advocated broad similarities between the Malmesbury/French Hoek/Klipheuwel sequence and the Pretoria/Rooiberg/Waterberg Beds of the Transvaal.

McIntyre's (1932) work on the central Cango showed a Pre-Cango Series of domomite, chert, granite and gneiss poorly exposed in several windows. The Cango Conglomerate, stripped of its separate status, was included as one of the units in the expanded stratigraphy included under the Cango Series. Lateral transitions between the quartzites and grits and the shales and limestones were attributed to sediment lensing.

The first and last systematic treatment of the pre-Cape rocks of the Gamtoos Valley in the eastern Cape was given by Amm (1934) and Frankel (1936), followed by Haughton and others, (1937b). They discerned a Lower pre-Cape stratigraphy consisting of two limestone units intercalated in phyllite and grit, followed by the Upper pre-Cape with a basal conglomerate and grit, passing upward through feldspathic quartzite and phyllite into the Table Mountain Sandstone. It is significant to note that this is the only place in the Cape where Cape sediments lie conformably upon rocks ascribed to the pre-Cape sequence. Frankel (1936) draws attention to similarities between these pre-Cape rocks and those of the Cango and Vanrhynsdorp area. Haughton and others (1937b) include them in the Malmesbury(?) Series and, as was customary at the time, in the Transvaal(?) System.

1940's

In 1946 Scholtz mentioned the various pre-Cape occurrences as supplementary information in his classic treatise on the younger intrusive granites of the Cape Province, in which he concluded that the granites of the south-western Cape were of two petrographically similar but distinct groups with similar age and that they appeared to be genetically related both to the George granite to the east and to the Kuboos Pluton in the Richtersveld. The granites of the Cape were thought to have been "emplaced towards the close of the Malmesbury orogenetic cycle under waning compressive stress acting from the W.S.W. or south-west." (Scholtz, 1946).

The locally legendary L. P. Rabie completed his detailed map of the Moorreesburg-Wellington area in 1948 and strode into obscurity, leaving behind no text to accompany this extraordinary piece of work. Two formations, the lower Swartland and the upper Boland, were distinguished in these Malmesbury beds. They are everywhere in fault contact and show a similar sequence of quartz and sericite schists interbedded with calcareous quartzite and grit. In addition, the beds at Heuningberg were mapped as younger material.

A map by persons at the University of Stellenbosch for the Worcester-Swellendam strip was also printed in 1948. It shows all of the pre-Cape sedimentary rocks as belonging to the Boland Formation which was in turn divided into an upper group of graywacke and conglomerate and a lower group of grit, quartzite, graywacke, and carbonate rock.

1950's

The year 1950 saw the publication of Potgieter's (1950b) dissertation on the George granite and intruded pre-Cape rocks of which he differentiated seven units composed of either quartzite or schist (Appendix C). Contact metamorphic effects largely obliterate the original sedimentary characteristics of these rocks and Potgieter (1950b) did not presume to make correlations with other pre-Cape rocks. The work is highlighted by a structural study which showed that the granite was intruded during deformation of the pre-Cape rocks and crystallized in a stress field directing pressure from S40W.

Two theses on parts of the Cango appeared in 1954. Stocken (1954) presented a stratigraphy scheme which was later closely followed on the Geological Survey map sheet of the area (Roussouw and others, 1964) (Appendix C). He also identified a unit of feldspathic grits unconformably overlying the Cango beds but underlying the Table Mountain Sandstone, which he correlated with the Klipheuwel of the western Cape. Clast studies demonstrated a provenance of sedimentary rocks similar to the Cango Formation with local granitic intrusions. Cross-bedding in the Cross-Bedded Grit Zone indicated a westerly source. Finally, Stocken (1954) correlated the Cango Formation with pre-Cape rocks in the Gamtoos and George areas. Mulder (1954) agreed with the Gamtoos correlation, but considered the pre-Cape rocks of George to bear little similarity to the Cango Formation in either composition or stratigraphy.

In a review of South African volcanism, Truter (1949) suggested that the Malmesbury might be equivalent to the rocks of the Gariep System in South West Africa, then thought to be Archean in age, and for the next decade the Malmesbury kept company with the ancients of the earth (de Toit, 1954; de Villiers, 1956; Visser, 1957; Haughton, 1963).

The Union map of 1955 showed a two-fold division of pre-Cape rock into Malmesbury and Klipheuwel, the second incorporating the former French Hoek beds as well as those of the former Malmesbury to the east of Rabie's (1948) supposed fault. The Klipheuwel was included in the Loskop System by the Survey whereas de Toit (1954) placed it in the Matsap, retaining the French Hoek as a separate group of rocks.

Data contributing to the Union Map of 1955 appeared when a mapping of the area between Worcester and Hermanus was published by de Villiers and others in 1964. The newly expanded Klipheuwel Formation which included previous French Hoek beds was thought to be of a separate, younger depositional and deformational cycle than the Malmesbury Formation.

Recent Work

Haughton (1963; 1969) and Visser (1967) both argued vigorously for the differentiation of the French Hoek from the Klipheuwel. Such a distinction was followed on the Republic map of 1970 (van Edden, 1970) with the Malmesbury, Cango and French Hoek beds included in the Nama System and the younger Klipeuwel correlated with nothing else in southern Africa.

Hartnady (1969) applied structural techniques to the rocks of the Worcester-French Hoek area and was able to distinguish early and late phases of deformation. The postulated Malmesbury-Klipheuwel unconformity north of Worcester was shown to be a fault and the stratigraphic order of the units there to be the reverse of the previous interpretation of de Villiers and others (1964).

Hartnady (1969) also concluded from their greatly differing structural states that the French Hoek and Klipheuwel Formations should not be correlated and suggested that certain sheared granites in the French Hoek area may represent crystalline basement upon which the French Hoek Formation was deposited. The Malmesbury of the area was renamed the Boland Group, in accord with Rabie's (1948) nomenclature for rocks farther to the west, with three formations in the Worcester area and two in areas to the south.

In a recent effort to provide a useful framework for future investigation, Hartnady and others (1974) proposed a new nomenclature for the Malmesbury Group in the southwestern Cape. Three tectonic domains separated by narrow belts of major dislocation were distinguished, and formations characteristic of each domain were defined (see Appendix C). The Tygerberg and Moorreesburg Formations are similar and consist of finely-bedded pelites and massively-bedded graywackes and immature quartzites, thought to be of turbidite origin. The phyllites, graywackes (without graded bedding), feldspathic grits and minor limestones of the Porterville and Piketburg Formations are interpreted as accumulating in a shelf environment.

In the southwestern tectonic domain tight, upright folding around northwest axes is characteristic, whereas folding is more complex in the central tectonic domain. Structures are again simpler in the northwestern tectonic domain around Porterville although in the Worcester area polyphase folding has been deciphered (Hartnady, 1969).

Cape Granites

The Cape Granites crop out as a group of sub-aligned, cross-cutting plutons primarily in the southwestern and central tectonic domains in the western Cape. It has been suggested by Scholtz (1946) that these are the discontinuous surface expression of a more extensive batholithic complex at depth. Kolbe (1966) showed with a detailed geochemical study that the Cape Granite formed from a high-level fractionated magma, with differentiation increasing strongly toward the outer margins of the plutons. Recent work by Visser and Schoch (1973) on the Darling and Vredenburg Plutons has distinguished multiple intrusive phases there.

Several isotopic age determinations on a pluton near Cape Town and on some Malmesbury rocks there provide the only age of the pre-Cape rocks in the Cape. A concordant U-Pb date of 610 ± 20 m.y. indicates the time of formation of the Cape Granite (Burger and Coertze, 1973) and gives a minimum age for deposition of the Malmesbury sediments that it intrudes. Homogenization of strontium in these rocks at 551 ± 8 m.y. is indicated by a Rb-Sr whole rock isochron (Allsopp and Kolbe, 1965). Cooling past the argon retention temperature for biotite occurred in the granite at 505 ± 25 m.y. (Aldrich and others, 1958).

Five whole-rock samples on the Malmesbury itself were co-linear giving a Rb-Sr isochron date of 595 ± 5 m.y. (Allsopp and Kolbe, 1965). This represents the time of strontium homogenization in the sediments and the authors conclude that deposition cannot greatly have exceeded this date.

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APPENDIX C: PRE-CAPE TABLES

An attempt to systematize the pre-Cape stratigraphy and correlations since their beginnings is represented by the Pre-Cape Tables. It is convenient to subdivide listings in the four main outcrop areas and assign to each a Roman numeral: I. Western Cape, II. Cango, III. George, and IV. Gamtoos. Authors and years of publication are listed chronologically according to usage of nomenclature, which is designated by letters: e.g., B. Malmesbury Clayslates, Beds, Series, Group. Correlations are cross-referenced by numerals and letters, with rock groups beyond the Cape indicated by Arabic numbers: e.g., 6. Waterberg. The symbol "=" means "correlated with" regardless of how much reservation was expressed by the author, "≠" means "not correlated with," and "<" means "younger than." Textbooks, review papers and regional maps are designated by "*".

I. SOUTHWESTERN CAPE

A. Clayslate and Gneiss
*Bain (1856) = IIA,IIIA,IVA

B. Malmesbury Clayslates, Beds, Series, Group
*Dunn (1887)
Rogers (1896)
Corstorphine (1897)
Rogers (1897)
Schwarz (1897a)
Corstorphine (1898)
Schwarz (1898a; 1898b)
Rogers and Schwarz (1898c)
Rogers (1903)
Schwarz (1905b) = 6a
*Hatch and Corstorphine (1905) = IIIC,8,12,14
*Rogers (1905)
Rogers, Schwarz and du Toit (1906a; 1906b)
Rogers and Schwarz (1907)
*Rogers and du Toit (1909) = IIIC,12,13
Rastall (1911)
Rogers, Schwarz and du Toit (1911)
Shand (1917)
*Rogers (1925) = IIC,1,3,9,10b
*du Toit (1926) = IIC,1b,1c,2a,9b,10b
*Schwarz (1928) = 1,9
*Rogers, Hall, Wagner and Haughton (1929) = 1,9
Haughton (1932) = 9a
Haughton (1933) = 9a
Söhnge (1934)
Beyers (1935)
*du Toit (1939) = IIC,1b,1c,2a,9b,10b
de Jager (1941)
*Scholtz (1946)
*Truter (1949) = 4,14
Potgeiter (1950a)
*du Toit (1954) = IIC,3,9b,10b
*Anonymous (1955) = IIC,4,14
de Villiers (1956)
*Visser (1957)
*Haughton (1963) = ID,2,4
de Villiers, Jansen and Mulder (1964) = 4
*Haughton (1969) = ID,IIC,2,4
*Truswell (1970)
*van Eeden (1970) = 1
Visser and Schoch (1973)
Hartnady, Newton and Theron (1974)

C. <u>Conglomerate at Honig Berg, Heuningberg Beds</u>

Schwarz (1898a) = IICa
*Rogers (1905)
*du Toit (1926) = IF,6a
*du Toit (1939) = IF, 6a
Rabie (1948)

D. <u>French Hoek Series, Beds</u>

Schwarz (1898b) = IICa
Rogers, Schwarz and du Toit (1906a)
*Rogers and du Toit (1909) = IE,IIC,5,11
Shand (1917)
*Rogers (1925) = IE,1,2b,9a,10a
*du Toit (1926) = IIC,IE,1a,2b,9a,10a
Haughton (1932) = 7
Haughton (1933) = 7
*du Toit (1939) = IICa,IE,1a,9a,10a
*Scholtz (1946)
*Truter (1949)
*du Toit (1954) = 2b,9a,10a
*Haughton (1963) = IB,2,4
Visser (1967) ≠IF
*Haughton (1969) = IB
Hartnady (1969) = IHb,e

E. <u>Ibiquas Series</u>

Rogers (1903)
Rogers, Schwarz and du Toit (1906b)
*Rogers and du Toit (1909) = ID,IIC,5,11
Rogers, Schwarz and du Toit (1911)
*Rogers (1925) = (see ID)

F. <u>Klipheuwel Series, Beds</u>

*du Toit (1926) = IC,6a
Haughton (1932) = IC,6
Haughton (1933) = IC,6
*du Toit (1939) = IC,6a
*du Toit (1954) = IC,6a
*Anonymous (1955) = 6b
*Haughton (1963) = 2a,2b
de Villiers, Jansen and Mulder (1964) = ID
Visser (1967) ≠ ID
*Haughton (1969)
*van Eeden (1970)
Hartnady, Newton and Theron (1974)

Ga. Boland Formation, Group
Rabie (1948)
Univ. of Stellenbosch (1948)
Hartnady (1969) = IIC
Hartnady, Newton and Theron (1974)

Gb. Swartland Formation
Rabie (1948)
Hartnady, Newton and Theron (1974)

Ha. Slanghoek Formation = IF
b. Waaihoek Formation = ID, IHe
c. Glen Heatlie Formation
d. Brandwacht Formation
e. Kaaimansgat Formation = IHb, ID
Hartnady (1969)

Ja. Brandwacht Formation
b. Porterville Formation
c. Piketberg Formation
d. Moorreesberg Formation
e. Porseleinberg Formation
f. Klipplaat Formation
g. Berg River Formation
h. Bridgetown Formation
i. Tygerberg Formation
Hartnady, Newton and Theron (1974)

II. CANGO VALLEY

A. Clayslate and Gneiss
*Bain (1856) = IA,IIIA,IVA

B. Namaqualand Schist
*Dunn (1887)

C. Cango Beds, Series, Formation
Corstorphine (1896a; 1896b)
Rogers and Schwarz (1897b)
Corstrophine (1898) ≠ IIB,<IB
Rogers and Schwart (1898)
*Hatch and Corstorphine (1905) = IE,IVB,10,11
*Rogers (1905) = IE
*Rogers and du Toit (1909) = ID,IE,5,11
*Rogers (1925) = IB (+ see IB)
*du Toit (1926) = IB (+ see IB)
*Schwarz (1928) = IB,1,9
*Rogers, Hall, Wagner and Haughton (1929) = IB,1,9
McIntyre (1932) = Vanrhynsdorp sediments
Rogers (1933)
*du Toit (1939) = IB (+ see IB)
*Scholtz (1946)
*Truter (1949) = IB (+ see IB)
King (1952)
*du Toit (1954) = IB (+ see IB)
Mulder (1954)
Stocken (1954) = IIID,4
*Anonymous (1955) = see IB)
Roussouw, Meyer, Mulder and Stocken (1964) = IVD
*Haughton (1969) = IB (+ see IB)
*Truswell (1970)
*van Eeden (1970) = 1

Ca. Cango Conglomerate, Series
Corstorphine (1898) <IB
Rogers and Schwarz (1898b)
*du Toit (1926) = ID (+ see ID)
*Schwarz (1928)
*du Toit (1939) = ID (+ see ID)
Mulder (1954)

Cb. Upper Fine-grained Series
Cango Conglomerate Series = IVDa
Lower Fine-grained Series = IVDb
Kango Valley Series
Mulder (1954)

Cc. <u>Unconformable Feldspathic Grit Formation</u>, <u>Klipheuwel Formation</u>

<u>Upper Graywacke Formation, Zone</u>
<u>Cross-bedded Grit Formation, Zone</u>
<u>Lower Graywacke Formation, Zone</u>
<u>Limestone (-Shale) Formation, Zone</u>
Stocken (1954) = IVDa and IVDb
Roussouw, Meyer, Mulder and Stocken (1964) = 4a and 4b
*Truswell (1970)

D. <u>Pre-Cango Series</u>

McIntyre (1932)
*Haughton (1963) = 13

III. GEORGE AREA

A. Clayslate and Gneiss
*Bain (1856) = IA,IIA,IVA

B. Namaqualand Schists
*Dunn (1887)

C. Malmesbury Beds
Corstorphine (1898) ≠ IIB
Rogers and Schwarz (1898a) distinct from IB
Rogers and Schwarz (1901)
Schwarz (1905a) hesitates to correlate with IB
*Hatch and Corstorphine (1905) = IB (+ see IB)
*Rogers (1905) = IB
*Rogers and du Toit (1909) = IB (+ see IB)
*Rogers (1925) = IB (+ see IB)
*du Toit (1926) = 13
*Schwarz (1928) = IB,1,9
*Rogers, Hall, Wagner and Haughton (1929) = IB,1,9
Haughton, Frommurze and Visser (1937a) = 9
*du Toit (1939) = IB,IIC
*Scholtz (1946)
*du Toit (1954) = IB,IIC (+ see IB)
*Anonymous (1955) = IB (+ see IB)
*van Eeden (1970) = 1

D. Homtini phyllites
Victoria Bay feldspathic quartzites
Victoria Bay phyllites
Kaaimansgat quartzites
Kaaimansgat phyllites
Groot Hoek quartz schist
Basal argillaceous horizon
Potgeiter (1950b)
*Haughton (1969)
*Truswell (1970)

IV. GAMTOOS VALLEY

A. <u>Clayslate and Gneiss</u>
*Bain (1856) = IA,IIA,IIIA

B. <u>Malmesbury Beds, Series</u>
*Rogers (1905) = IB
*Rogers and du Toit (1909) - IB (+ see IB)
*Schwarz (1928) = IB,1,9
Haughton (1928)
*Rogers, Hall, Wagner and Haughton (1929) = IB,1,9

C. <u>Cango Series</u>
*Hatch and Corstorphine (1905) = IIC (+ see IIC)
*Rogers (1925) = IB (+ see IB)
*du Toit (1926) = IIC (+ see IB)

D. <u>Pre-Cape Series</u>
Amm (1934) = IB
Frankel (1936) = IIC,IICa
Haughton, Frommurze and Visser (1937b) = IB,9
*du Toit (1939) = IIC (+ see IB)
*Scholtz (1946)
*du Toit (1954) = IIC (+ see IB)
*Haughton (1969)
*van Eeden (1970) = 1

Da. <u>Upper Pre-Cape</u>
Frankel (1936) = Ibiquas (Vanrhynsdorp area)
Haughton, Frommurze and Visser (1937b)
*Anonymous (1955) = IF (+ see IF)
*Haughton (1969)

Db. <u>Lower Pre-Cape</u>
Frankel (1936 = Malmesbury (Vanrhynsdorp area)
Haughton, Frommurze and Visser (1937b)
*Anonymous (1955) = IB (+ see IB)
*Haughton (1969)

Correlated Rock Groups Outside of the Cape

1. Nama	1a. Fish River
	1b. Schwarzrand
	1c. Schwarzkalk
	1d. Kuibis
2. Damara	2a. Otavi dolomite
	2b. Numees
3. Kaigas	
4. Gariep	4a. Holgat
	4b. Hilda
5. Koras	
6. Waterberg	
6a. Matsap	
6b. Loskop	
7. Rooiberg	
8. Namaqualand	
9. Transvaal	9a. Pretoria
	9b. Dolomite
	9c. Black Reef
10. Griqualand West	10a. Griquatown
	10b. Campbell Rand
	10c. Black Reef
11. Ventersdorp	
12. Swaziland	
13. Kheis	
14. Archean	

REFERENCES

Aaron, J.M., and Ford, A.B. 1964. Isotopic age determinations in the Thiel Mountains, Antarctica. Geol. Soc. America Spec. Paper 76, Abstracts for 1963, p. 1.

Aldrich, L.T., Wetherill, G.W., Davis, G.L., and Tilton, G.R. 1958. Radioactive ages of micas from granitic rocks by Rb-Sr and K-A methods. Am. Geophys. Trans., v. 39, p. 1124-1134.

Allsopp, H.L., and Kolbe, P. 1965. Isotopic age determinations on the Cape Granite and intruded Malmesbury sediments, Cape Peninsula, South Africa. Geochim. et Cosmochim. Acta, v. 29, p. 1115-1130.

Amm, E.L. 1934. The pre-Cape rocks of the Uitenhage division. Geol. Soc. South Africa Trans., v. 37, p. 69-86.

Anonymous. 1955. Geological map of the Union of South Africa. South Africa Geol. Survey.

Armstrong, R.L. 1967. K-Ar dating of plutonic and volcanic rocks in orogenic belts: in Schaeffer, O.A., and Zahringer, J., eds., Potassium Argon Dating, Springer-Verlag, New York, p. 117-133.

Bain, A.G. 1856. On the geology of South Africa. Geol. Soc. London Trans., 2nd Series, v. 7, p. 175-192.

Balk, R. 1937. Structural behavior of igneous rocks. Geol. Soc. America Mem. 5. 177 p.

Barrett, P.J. 1965. Geology of the area between the Axel Heiberg and Shackleton Glaciers, Queen Maud Range, Antarctica, Part 2 - Beacon Group. New Zealand Jour. Geology and Geophysics, v. 8, p. 344-370.

Barrett, P.J., and Elliot, D.H. 1973. Reconnaissance geologic map of the Buckley Island Quadrangle, Transantarctic Mountains, Antarctica. Antarctic Geol. Map A-3, U.S. Ant. Res. Program.

Barrett, P.J., Grindley, G.W., and Webb, P.N. 1972. The Beacon Supergroup of East Antarctica: in Adie, R.J., ed., Antarctic Geology and Geophysics, Universitetsforlaget, Oslo, p. 319-332.

Barrett, P.J., Lindsay, J.F., and Gunner, J. 1970. Reconnaissance geologic map of the Mount Rabot Quadrangle, Transantarctic Mountains, Antarctica. Antarctic Geol. Map A-1, U.S. Ant. Res. Program.

Beyers, H.P. 1935. Geology of an area south of the Rivier Zonder Einde Range. Geol. Soc. South Africa Trans., v. 38, p. 91-92.

Blackburn, Q.A. 1937. The Thorne Glacier section of the Queen Maud Mountains. Geog. Rev., v. 27, p. 598-614.

Bouma, A.H. 1962. Sedimentology of some flysch deposits, a graphic approach to facies interpretation. Elsevier, Amsterdam, 168 p.

Brink, W.C. 1950. The geology, structure and petrology of the Nuwerus area. Univ. Stellenbosch Annals, v. 26(A), p. 97-221.

Bührmann, H.T. 1969. The geology of the lower Kobe Valley, Vanrhynsdorp district. M.Sc. Thesis, Univ. Stellenbosch. 103 p.

Burgener, J.D. 1975. Petrography of the Queen Maud Batholith, Central Transantarctic Mountains, Ross Dependency, Antarctica. M.Sc. Thesis, Univ. Wisconsin. 75 p.

Burger, A.J., and Coertze, F.J. 1973. Radiometric age measurements on rocks from southern Africa to the end of 1971. South Africa Geol. Survey Bull. 58, 46 p.

Clifford, T.N. 1967. The Damaran episode in the upper Proterozoic-lower Paleozoic structural history of southern Africa. Geol. Soc. America Spec. Paper 92, 100 p.

Cook, H.E., McDaniel, P.N., Mountjoy, E.W., and Pray, L.D. 1972. Allochthonous carbonate debris flows at Devonian bank ('reef') margins, Alberta, Canada. Bull. Canadian Petroleum Geology, v. 20, p. 439-497.

Corstorphine, G.S. 1896a. Report of a preliminary geological survey of the Oudtshoorn and Prince Albert districts with a view to the selection of a site for a deep borehole for artesian waters. Cape of Good Hope, Geol. Comm. Ann. Rept. 1, p. 31-33.

Corstorphine, G.S. 1896b. The Cango Cave. Cape of Good Hope, Geol. Comm. Ann. Rept. 1, p. 34-36.

Corstorphine, G.S. 1897. Geologist's report for 1897. Cape of Good Hope, Geol. Comm. Ann. Rept. 2, p. 3-43.

Corstorphine, G.S. 1898. Geologist's report for 1898. Cape of Good Hope, Geol. Comm. Ann. Rept. 3, p. 3-13.

Craddock, C. 1969. Geology of the Ellsworth Mountains. Am. Geog. Soc., Antarctic Map Folio Ser. 12, IV.

Craddock, C., Anderson, J.J., and Webers, G.F. 1964. Geologic outline of the Ellsworth Mountains: in Adie, R.J., ed., Antarctic Geology, North-Holland, Amsterdam, p. 155-170.

David, T.W.E. and Priestly, R.E. 1914. Glaciology, physiography, stratigraphy, and tectonic geology of South Victoria Land, with short notes on paleontology by T.G. Taylor and E.G. Goddard. Brit. Ant. Exped., 1907-09, Geol., v. 1, Heinemann, London. 319 p.

Debenham, F. 1921. The sandstone, etc., of the McMurdo Sound, Terra Nova Bay, Beardmore Glacier regions. Brit. Ant. ('Terra Nova') Exped., 1910. Nat. Hist. Rept., Geol., v. 1, no. 4a, Brit. Mus. (Nat. Hist.), London, p. 103-119.

de Jager, F.S.J. 1941. The Malmesbury succession east of the Robertson granite. M.Sc. Thesis, Univ. Stellenbosch.

de Villiers, J. 1956. Die drie sintaksisse in die suidwestelike Kaapprovinsie. Tegnikon, v. 9, p. 76-87.

de Villiers, J., and Burger, A.J. 1967. Note on the minimum age of certain granites from the Richtersveld area. South Africa Geol. Survey Annals, v. 6, p. 83-84.

de Villiers, J., Jansen, H., and Mulder, M.P. 1964. Die geologie van die gebied tussen Worcester en Hermanus. Toeligting van gebiede 3319C (Worcester) en 3419A (Caledon) en dele van gebiede 3318D (Stellenbosch) en 3418B (Somerset-Wes), South Africa Geol. Survey. 69 p.

de Villiers, J., and Söhnge, P.G. 1959. The geology of the Richtersveld. South Africa Geol. Survey Mem. 48. 295 p.

Doumani, G.A. and Minshew, V.H. 1965. General geology of the Mount Weaver area, Queen Maud Mountains, Antarctica. Am. Geophys. Union, Antarctic Res. Ser., v. 6, p. 127-139.

Dunn, E.J. 1973. Geological sketch map of Cape Colony: Sand Cast, Stanford, London.

du Toit, A.L. 1926. The geology of South Africa, 1st ed. Oliver and Boyd, Edinburg and London. 463 p.

du Toit, A.L. 1939. The Geology of South Africa, 2nd ed. Oliver and Boyd, Edinburgh and London. 539 p.

du Toit, A.L. 1954. The geology of South Africa, 3rd ed. Oliver and Boyd, Edinburgh and London. 611 p.

Eastin, R. 1970. Geochronology of the basement rocks of the central Transantarctic Mountains, Antarctica. Ph.D. Dissertation, The Ohio State Univ. 215 p.

Eastin, R., and Faure, G. 1971. The age of the Littlewood volcanics of Coats Land, Antarctica. Jour. Geol. v. 79, p. 241-245.

Eastin, R., and Faure, G. 1972. Geochronology of the basement rocks of the Pensacola Mountains, Antarctica. Geol. Soc. America Absts. with Programs, v. 4, p. 496.

Elliot, D.H. 1972. Aspects of Antarctic geology and drift reconstructions: in Adie, R.J., ed., Antarctic Geology and Geophysics, Universitetsforlaget, Oslo, p. 849-857.

Elliot, D.H., Barrett, P.J., and Mayewski, P.A. 1974. Reconnaissance geologic map of the Plunket Point Quadrangle, Transantarctic Mountains, Antarctica. Antarctic Geol. Map A-4, U.S. Ant. Res. Program.

Faure, G., Murtaugh, J.G., and Montigny, R.J.E. 1968. The geology and geochronology of the basement complex of the central Transantarctic Mountains. Can. Jour. Earth Sci., v. 5, p. 555-560.

Ferrar, H.T. 1907. Report on the field geology of the region explored during the 'Discovery' Antarctic Expedition, 1901-04: Nat. Ant. Exped., 1901-04. Nat. Hist., Geol., v. 1, Brit. Mus. (Nat. Hist.), London. p. 1-100.

Flores, G. 1970. Suggested origin of the Mozambique Channel. Geol. Soc. South Africa Trans., v. 73, p. 1-17.

Ford, A.B. 1964. Cordierite-bearing, hypersthene-quartz-monzonite porphyry in the Thiel Mountains, and its regional importance: in Adie, R.J., ed., Antarctic Geology, North-Holland, Amsterdam, p. 429-441.

Ford, A.B. and Aaron, J.M. 1962. Bedrock geology of the Thiel Mountains, Antarctica. Science, v. 137, p. 751-752.

Ford, A.B., Hubbard, H.A., and Stern, T.W. 1963. Lead-alpha ages of zircon in quartz monzonite porphyry, Thiel Mountains, Antarctica. A preliminary report. U.S. Geol. Survey Prof. Paper 450-E, p. 105-107.

Ford, A.B., and Sumsion, R.S. 1971. Late Precambrian silicic pyroclastic volcanism in the Thiel Mountains. Antarctic Jour. U.S., v. 6, p. 185-186.

Frakes, L.A., and Crowell, J.C. 1968. Late Paleozoic glacial facies and the origin of the South Atlantic basin. Nature, v. 217, p. 837-838.

Frankel, J.J. 1936. Geology of a portion of the Gamtoos Valley. Geol. Soc. South Africa Trans., v. 39, p. 263-279.

Frets, D.C. 1969. Geology and structure of the Haub-Welwitchia area, South West Africa. Precambrian Res. Unit. Bull. 5, Univ. Cape Town, 235 p.

Gair, H.S., Sturm, A., Carryer, S.J., and Grindley, G.W. 1969. The geology of northern Victoria Land. Am. Geog. Soc., Antarctic Map Folio Ser. 12, XII.

Germs, G.J.B. 1972a. The stratigraphy and paleontology of the lower Nama Group, South West Africa. Precambrian Res. Unit. Bull. 12, Univ. Cape Town. 250 p.

Germs, G.J.B. 1972b. New shelly fossils from the Nama Group, South West Africa. Am. Jour. Sci., v. 272, p. 752-761.

Germs, G.J.B. 1974. The Name Group in South West Africa and its relationship to the Pan-African geosyncline. Jour. Geol., v. 82, p. 301-317.

Gevers, T.W. 1931. An ancient tillite in South West Africa. Geol. Soc. South Africa Trans., v. 34, p. 1-17.

Gevers, T.W. 1934. The geology of the Windhoek district of South West Africa. Geol. Soc. South Africa Trans., v. 37, p. 221-251.

Gould, L.M. 1931. Some geographical results of the Byrd Antarctic expedition. Geog. Rev., v. 21, p 177-200.

Gould, L.M. 1935. Structure of the Queen Maud Mountains. Geol. Soc. America Bull., v. 46, p. 973-984.

Grindley, G.W. 1963. The geology of the Queen Alexandra Range, Beardmore Glacier, Ross Dependency, Antarctica; with notes on the correlation of Gondwana sequences. New Zealand Jour. Geology and Geophysics, v. 6, p. 307-347.

Grindley, G.W. 1972. Polyphase deformation of the Precambrian Nimrod Group, central Transantarctic Mountains: in Adie, R.J., ed., Antarctic Geology and Geophysics, Universitetsforlaget, Oslo, p. 313-318.

Grindley, G.W., and Laird, M.G. 1969. Geology of the Shackleton Coast. Am. Geog. Soc., Antarctic Map Folio Ser. 12, XIV.

Grindley, G.W. and McDougall, I. 1969. Age and correlation of the Nimrod Group and other Precambrian rock units in the central Transantarctic Mountains, Antarctica. New Zealand Jour. Geology and Geophys, v. 12, p. 391-411.

Grindley, G.W., McGregor, V.R., and Walcott, R.I. 1964. Outline of the geology of the Nimrod-Beardmore-Axel Heiberg Glaciers region, Ross Dependency: in Adie, R.J., Antarctic Geology, North-Holland, Amsterdam, p. 206-219.

Guj, P. 1967. Structural geology of the Auas Mountains, Windhoek district, South West Africa. South Africa Geol. Survey Annals, v. 6, p. 55-62.

Guj, P. 1970. The Damara mobile belt in the south-western Kaokoveld. Precambrian Res. Unit. Bull. 8, Univ. Cape Town. 168 p.

Gunn, G.M., and Walcott, R.I. 1962. The geology of the Mt. Markam region, Ross Dependency, Antarctica. New Zealand Jour. Geology and Geophysics, v. 5, p. 407-426.

Gunn, G.M. and Warren, G. 1962. Geology of Victoria Land between Mawson and Mullock Glaciers, Antarctica. New Zealand Geol. Survey Bull. 71. 157 p.

Gunner, J. 1969. Petrography of metamorphic rocks from the Miller Range, Antarctica. Inst. Polar Stud. Rept. 32, The Ohio State Univ. 44 p.

Gunner, J. 1971a. Age and origin of the Nimrod Group and of the Granite Harbour Intrusives, Beardmore Glacier region, Antarctica. Ph.D. Dissertation, Ohio State Univ. 231 p.

Gunner, J. 1971b. Ida Granite: a new formation of the Granite Harbour Intrusives, Beardmore Glacier region. Antarctic Jour. U.S., v. 6, p. 194-196.

Gunner, J. 1974. Investigations of lower Paleozoic granites in the Beardmore Glacier region. Antarctic Jour. U.S., v. 9, p. 76-81.

Gunner, J., and Faure, G. 1972. Rubidium-strontium geochronology of the Nimrod Group, central Transantarctic Mountains: in Adie, R.J., ed., Antarctic Geology and Geophysics, Universitetsforlaget, Oslo, p. 305-311.

Gunner, J., and Mattinson, J.M. 1975. Rb-Sr and U-Pb isotopic ages on granites in the central Transantarctic Mountains. Geol. Mag., v. 112, p. 25-31.

Haller, J. 1955. Der "zentral metamorphe komplex" von Nordostgrønland. Teil I. Die geologische karte von Suess Land, Gletscherland und Goodenoughs Land. Grønland, v. 73(I), p. 1-174.

Haller, J. 1971. Geology of the East Greenland Caledonides. Wiley-Interscience, New York. 413 p.

Hamilton, W. 1960. New interpretation of Antarctic tectonics. U.S. Geol. Survey Prof. Paper 400-B, p. B379-B380.

Hamilton, W. 1967. Tectonics of Antarctica. Tectonophysics, v. 4, p. 555-568.

Harrignton, H.J. 1958. Nomenclature of rock units in the Ross Sea region, Antarctica. Nature, v. 182, p. 290.

Harrington, H.J., Wood, B.L., McKellar, I.C., and Lensen, G.J. 1967. Topography and geology of the Cape Hallet district, Victoria Land, Antarctica. New Zealand Geol. Survey Bull. 80. 100 p.

Hartnady, C.J. 1969. Structural analysis of some pre-Cape formations in the western province. Precambrian Res. Unit. Bull. 6, Univ. Cape Town. 70 p.

Hartnady, C.J., Newton, A.R., and Theron, J.N. 1974. The stratigraphy and structure of the Malmesbury Group in the southwestern Cape. Precambrian Res. Unit. Bull. 15, Univ. Cape Town, p. 193-213.

Haskell, T.R., Kennett, J.P., and Prebble, W.M., 1965. Geology of the Brown Hills and Darwin Mountains, Southern Victoria Land, Antarctica. Transactions, Royal Soc. of New Zealand, Geology, v. II, p.231-248.

Hatch, F.H., and Corstorphine, G.S. 1905. The geology of South Africa. McMillan, London. 348 p.

Haughton, S.H. 1928. The geology of the country between Grahamstown and Port Elizabeth: An explanation of Sheet 9 (Post Elizabeth). South Africa Geol. Survey. 45 p.

Haughton, S.H. 1932. The pre-Cape succession in the western province: Geol. Soc. South Africa Trans., v. 35, p. 193-196.

Haughton, S.H. 1933. The geology of Cape Town and adjoining country: An explanation of Sheet 247 (Cape Town). South Africa Geol. Survey. 90 p.

Haughton, S.H. 1963. The stratigraphic history of Africa south of the Sahara. Oliver and Boyd, Edinburgh and London. 365 p.

Haughton, S.H., Frommurze, H.F., and Visser, D.J.L. 1937a. The geology of the country around Mossel Bay, Cape Province: An explanation of Sheet 201 (Mossel Bay). South Africa Geol. Survey. 48 p.

Haughton, S.H., Frommurze, H.F., and Visser, D.J.L. 1937b. The geology of a portion of the coastal belt near the Gamtoos Valley, Cape Province: An explanation of Sheets 151 North and 151 South (Gamtoos River). South Africa Geol. Survey. 55 p.

Hill, D. 1964. Archaeocyatha from the Shackleton Limestone of the Ross System, Nimrod Glacier area, Antarctica. Royal Soc. New Zealand Trans., Geology, v. 2, p. 137-146.

Hill, D. 1965. Archaeocyatha from Antarctica and a review of the phylum. Trans-Antarctic Expedition Sci. Rept. 10. 151 p.

Irvine, T.N. and Baragar, W.R.A. 1971. A guide to the chemical classification of the common volcanic rocks. Canadian Jour. Earth Sciences, v. 8, p. 523-548.

Jansen, H. 1960. The geology of the Bitterfontein area, Cape Province: An explanation of Sheet 253 (Bitterfontein). South Africa Geol. Survey.

Kanmera, K. 1974. Paleozoic and Mesozoic geosynclinal volcanism in the Japanese Islands and associated chert sedimentation: in Dott, R.H., Jr., and Shaver, R.H., eds., Modern and Ancient Geosynclinal Sedimentation, Soc. Econ. Paleontologists and Mineralogists Spec. Pub. 19, p. 161-173.

Katz, H.R. and Waterhouse, G.C. 1970a. Geological reconnaissance of the Scott Glacier area, south-eastern Queen Maud Range, Antarctica. New Zealand Jour. Geology and Geophysics, v. 13, p. 1030-1037.

Katz, H.R., and Waterhouse, G.C. 1970b. The geologic situation at O'Brian Peak, Queen Maud Range, Antarctica. New Zealand Jour. Geology and Geophysics, v. 13, p. 1038-1047.

Kennedy, W.Q. 1964. The structural differentiation of Africa in the Pan African (± 500 m.y.) tectonic episide. Res. Inst. Afr. Geology Ann. Rept. (1962-63), v. 8, Univ. Leeds, p. 48-49.

Kolbe, P. 1966. Geochemical investigation of the Cape Granite, south-western Cape Province, South Africa. Geol. Soc. South Africa Trans., v. 69, p. 161-199.

Korn, H., and Martin, H. 1959. Gravity tectonics in the Naukluft Mountains of South West Africa. Geol. Soc. America Bull., v. 70, p. 1047-1078.

Kröner, A. 1968. The gneiss-sediment relationships north-west of Vanrhynsdorp, Cape Province. Precambrian Res. Unit Bull. 3, Univ. Cape Town. 233 p.

Kröner, A. 1969. The correlation of the pre-Cape sediments in the Vanrhynsdorp region, Cape Province. Geol. Soc. South Africa Trans., v. 72, p. 127-150.

Kröner, A. 1971. Late-Precambrian correlation and the relationship between the Damara and Nama Systems of South West Africa. Geol. Rundschau, v. 60, p. 1513-1523.

Kröner, A. 1974. Editorial comments concerning the present confusion over lithostratigraphic nomenclature and a proposal for a new Damara terminology. Precambrian Res. Unit Bull. 15, Univ. Cape Town, p. 139-140.

Kröner, A., and Germs, G.J.B. 1971. A re-interpretation of the Numees-Nama contact at Aussenkjer, South West Africa. Geol. Soc. South Africa Trans., v. 74, p. 69-74.

Kröner, A. and Rankama, K. 1972. Late Precambrian glaciogenic sedimentary rocks in southern Africa: a compilation with definitions and correlations. Precambrian Res. Unit Bull. 11, Univ. Cape Town. 37 p.

Laird, M.G. 1963. Geomorphology and stratigraphy of the Nimrod Glacier-Beaumont Bay region, southern Victoria Land, Antarctica. New Zealand Jour. Geology and Geophysics, v. 6, p. 465-484.

Laird, M.G. 1964. Petrology of rocks from the Nimrod Glacier-Starshot Glacier region, Ross Dependency: in Adie, R.J., ed., Antarctic Geology, North-Holland, Amsterdam, p. 463-472.

Laird, M.G., Mansergh, G.D., and Chappell, J.M.A. 1971. Geology of the central Nimrod Glacier area, Antarctica. New Zealand Jour. Geology and Geophysics, v. 14, p. 427-468.

Laird, M.G., and Waterhouse, J.B. 1962. Archeocyathine limestones of Antarctica. Nature, v. 194, p. 861.

Lamont, G.T. 1947. The geology of part of the Vanrhynsdorp division, Cape Province. Ph.D. Dissertation, Univ. Cape Town.

LaPrade, K.E. 1969. Geology of Shackleton Glacier area, Queen Maud Range, Transantarctic Mountains, Antarctica. Ph.D. Dissertation. Texas Tech. Univ.

Linder, H., de Hills, S.M., and Thiel, E.C. 1965. Basement complex in the Queen Maud Mountains, central Transantarctic Mountains. Am. Geophys. Union, Antarctic Res. Ser., v. 6, p. 141-144.

Lindsay, J.F., Gunner, J., and Barrett, P.J. 1973. Reconnaissance geologic map of the Mount Elizabeth and Mount Kathleen Quadrangles, Transantarctic Mountains, Antarctica. Antarctic Geol. Map A-2, U.S. Ant. Res. Program.

Lowe, D.R., and LoPiccolo, R.D. 1974. The characteristics and origins of dish and pillar structures. Jour. Sed. Petrology, v. 44, p. 484-501.

Martin, H. 1965. The Precambrian geology of South West Africa and Namaqualand. Precambrian Res. Unit, Univ. Cape Town. 159 p.

Martin, H. 1974. Damara rocks as nappes on the Naukluft Mountains, South West Africa. Precambrian Res. Unit Bull. 15, Univ. Cape Town, p. 153-165.

Mawson, D. 1916. Petrology of rock collections from the mainland of South Victoria Land. Brit. Ant. Exped., 1907-09. Geol., v. 2, Heinemann, London, p. 201-234.

McDougall, I., and Grindley, G.W. 1965. Potassium-argon dates on micas from the Nimrod-Beardmore-Axel Heiberg region, Ross Dependency, Antarctica. New Zealand Jour. Geology and Geophysics, v. 8, p. 304-313.

McElhinny, M.W., Embleton, B.J.J., Daly, L. and J-P. Pozzi. 1976. Paleomagnetic evidence for the location of Madagascar in Gondwanaland. Geology, v. 4, p. 455-457.

McGregor, V.R. 1965. Geology of the area between the Axel-Heiberg and Shackleton Glaciers, Queen Maud Range, Antarctica. New Zealand Jour. Geology and Geophysics, v. 8, p. 314-343.

McGregor, V.R., and Wade, F.A. 1969. Geology of the western Queen Maud Mountains. Am. Geog. Soc. Map Folio Ser. 12, XV.

McIntyre, R.D. 1932. The geology of the Cango beds. Geol. Soc. South Africa Trans., v. 35, p. 69-84.

McLelland, D. Unpub. Geology of the basement complex, Nilsen Plateau, Antarctica. Unpub. ms.

Middleton, G.V., and Hampton, M.A. 1973. Sedimentary gravity flows; mechanics of flow and deposition. Soc. Econ. Paleontologists and Mineralogists, Short Course, Anaheim, p. 1-38.

Minshew, V.H. 1965. Potassium-argon age from a granite at Mount Wilber, Queen Maud Range, Antarctica. Science, v. 150, p. 741-743.

Minshew, V.H. 1966. Stratigraphy of the Wisconsin Range, Horlick Mountains, Antarctica. Science, v. 152, p. 37-38.

Minshew, V.H. 1967. Geology of the Scott Glacier and Wisconsin Range areas, central Transantarctic Mountains, Antarctic. Ph.D. Dissertation, The Ohio State Univ. 268 p.

Mirsky, A. 1969. Geology of the Ohio Range-Liv Glacier area. Am. Geog. Soc. Map Folio Ser. 12, XVII.

Moore, J.G., and Peck, D.L. 1962. Accretionary lapilli in volcanic rocks of the western continental United States. Jour. Geol., v. 70, p. 182-193.

Mulder, M.P. 1954. The geology of the western Kango. M.Sc. Thesis, Univ. Witwatersrand.

Murtaugh, J.G. 1969. Geology of the Wisconsin Range Batholith, Transantarctic Mountains. New Zealand Jour. Geology and Geophysics, v. 12, p. 526-550.

Nelson, W.H., Schmidt, D.L., and Schopf, J.M. 1969. Structure and stratigraphy of the Pensacola Mountains, Antarctica. Geol. Soc. America Spec. Paper 115, Abstracts for 1967, p. 344.

Oliver, R.L. 1964. Some basement rock relations in Antarctica: in Adie, R.J., ed., Antarctic Geology, North-Holland, Amsterdam, p. 259-264.

Oliver, R.L. 1972. The geology of an area near the mouth of the Beardmore Glacier, Ross Dependency, Antarctica: in Adie, R.J., ed., Antarctic Geology and Geophysics, Universitetsforlaget, Oslo, p. 379-385.

Palmer, A.R. and Gatehouse, C.G. 1972. Early and middle Cambrian trilobites from Antarctica. U.S. Geol. Survey Prof. Paper 456-D. 37 p.

Pettijohn, F.J., Potter, P.E., and Siever, R. 1972. Sand and sandstone. Springer-Verlag, New York. 618 p.

Potgieter, C.T. 1950a. The South African pyrophyllite deposits. Univ. Stellenbosch Annals, v. 26(A), p.244-265.

Potgieter, C.T. 1950b. The structure and petrology of the George Granite plutons and the invaded pre-Cape sedimentary rocks. Univ. Stellenbosch Annals, v. 26(A), p. 323-412.

Rabie, L.P. 1948. Geological map of the Morreesburg-Wellington area. Unpub. Univ. Stellenbosch.

Rastall, R.H. 1911. The geology of the districts of Worcester, Robertson and Ashton (Cape Coloney). Geol. Soc. London Quart. Jour., v. 67, p. 701-733.

Robinson, E.S. 1964. Geological structure of the Transantarctic Mountains and adjacent ice covered areas, Antarctica. Ph.D. Dissertation. Univ. Wisconsin. 290 p.

Rodgers, J. 1970. The tectonics of the Appalachians. Wiley-Interscience, New York. 271 p.

Rogers, A.W. 1896. Summary of the work in the south-western districts, Cape of Good Hope, Geol. Comm. Ann. Rept. 1, p. 13-14.

Rogers, A.W. 1897. Survey of the Stellenbosch district. Cape of Good Hope. Geol. Comm. Ann. Rept. 2, p. 47-49.

Rogers, A.W. 1903. Geological survey of the divisions of Piquetberg, Clanwilliam, and Van Rhyn's Dorp. Cape of Good Hope, Geol. Comm. Ann. Rept. 8, p. 139-167.

Rogers, A.W. 1904. Geological survey of the north-western part of Van Rhyn's Dorp: Cape of Good Hope, Geol. Comm. Ann. Rept. 9, p. 9-46.

Rogers, A.W. 1905. An introduction to the geology of Cape Coloney. Green, London.

Rogers, A.W. 1911. Geological survey of parts of Van Rhynsdorp and Namaqualand divisions. Cape of Good Hope, Geol. Comm. Ann. Rept. 14, p. 7-84.

Rogers, A.W. 1912. The Nama System in the Cape Province. Geol. Soc. South Africa Trans., v. 15, p. 31-50.

Rogers, A.W. 1925. The geological structure of the Union: An explanation of the geological map of the Union of South Africa, on the scale of one to a million. South Africa Geol. Survey. 34 p.

Rogers, A.W. 1933. Discussion on the paper by R.C. McIntyre: Cango beds. Geol. Soc. South Africa Proc., v. 35, p. 41.

Rogers, A.W. and du Toit, A.L. 1909. An introduction to the geology of Cape Coloney. Longmans, Green & Co., London. 491 p.

Rogers, W.W., Hall, A.L., Wagner, P.A. and Haughton, S.H. 1929. The Union of South Africa. Handbuch der regionalen geologie, VII Band, 7a abteilung. 232 p.

Rogers, A.W., and Schwarz, E.H.L. 1897. Summary of work done during 1897 between the Karroo and the Langebergen. Cape of Good Hope, Geol. Comm. Ann. Rept. 2, p. 56-69.

Rogers, A.W., and Schwarz, E.H.L. 1898a. Report on the southern districts between Breede River and George. Cape of Good Hope, Geol. Comm. Ann. Rept. 3, p. 55-64.

Rogers, A.W., and Schwarz, E.H.L. 1898b. Report on Oudtshoorn. Cape of Good Hope, Geol. Comm. Ann. Rept. 3, p. 65-79.

Rogers, A.W. and Schwarz, E.H.L. 1898c. Notes on the geology round Worcester supplementary to the accounts published in the first and second annual reports. Cape of Good Hope, Geol. Comm. Ann. Rept. 3, p. 81-84.

Rogers, A.W., and Schwarz, E.H.L. 1900. Report on the survey of parts of the Clanwilliam, Van Rhyn's Dorp, and Calvinia divisions. Cape of Good Hope, Geol. Comm. Ann. Rept. 5, p. 19-54.

Rogers, A.W., and Schwarz, E.H.L. 1901. Report on a journey from Swellendam to Mossel Bay. Cape of Good Hope, Geol. Comm. Ann. Rept. 6, p. 7-21.

Rogers, A.W., and Schwarz, E.H.L. 1907. Geological map of the Coloney of the Cape of Good Hope, Sheet II. Cape of Good Hope, Geol. Comm.

Rogers, A.W., Schwarz, E.H.L., and du Toit, A.L. 1906a. Geological map of the Coloney of the Cape of Good Hope, Sheet I. Cape of Good Hope, Geol. Comm.

Rogers, A.W., Schwarz, E.H.L. and du Toit, A.L. 1906b. Geological map of the Coloney of the Cape of Good Hope, Sheet IV. Cape of Good Hope, Geol. Comm.

Rogers, A.W., Schwarz, E.H.L., and du Toit, A.L. 1911. Geological map of the Province of the Cape of Good Hope, Sheet XI. Cape of Good Hope, Geol. Comm.

Rose, W.I., Jr. 1972. Santiaguito volcanic dome, Guatemala. Geological Soc. of America Bull., v. 83, p. 1413-1434.

Ross, C.S., and Smith, R.L. 1961. Ash flow tuffs: their origin, geologic relations and identification. U.S. Geol. Survey Prof. Paper 366. 81 p.

Ross, J.C. 1847. A voyage of discovery and research in the southern and Antarctic regions during the years 1839-43. Murray, London, v. 1, 366 p., v. 2, 447 p.

Roussouw, P.J., Meyer, E.I., Mulder, M.P., Stocken, C.G. 1946. Die geologie van die Swartberge, die Kangovallei en die ongewing van Prins Albert, Kapprovinsie. Toelightig van Blaaie 3321B en 3322A, South Africa Geol. Survey. 93 p.

Rubridge, R.N. 1859. On some points in the geology of South Africa. Quart. Jour. Geol. Soc., v. 15, p. 195-198.

Sanders, J.E. 1965. Primary sedimentary structures formed by turbidity currents and related resedimentation mechanisms. Soc. Econ. Paleontologists and Mineralogists Spec. Pub. 12, p. 192-219.

Schalk, K.E.L. 1970. Some late Precambrian formations in central South-West Africa. South Africa Geol. Survey Annals, v. 8, p. 29-48.

Schetelig, J. 1915. Report on the rock specimens collected on Roald Amundsen's South Pole expedition. Skr. Vidensk Selsk. Krist., I. Mat.-Naturv., no. 4, p. 1-32.

Schmidt, D.L., Dover, J.H., Ford, A.B., and Brown, R.D. 1964. Geology of the Patuxent Mountains: in Adie, R.J., ed., Antarctic Geology, North-Holland, Amsterdam, p. 276-283.

Schmidt, D.L., and Ford, A.B. 1969. Geology of the Pensacola and Thiel Mountains. Am. Geog. Soc. Map Folio Ser. 12, V.

Schmidt, D.L., Williams, P.L., Nelson, W.H., and Ege, J.R. 1965. Upper Precambrian and Paleozoic stratigraphy and structure of the Neptune Range, Antarctica. U.S. Geol. Survey Prof. Paper 525-D, p. D112-D119.

Scholtz, D.L. 1946. On the younger Pre-Cambrian granite plutons of the Cape Province. Geol. Soc. South Africa Proc., v. 49, p. 35-82.

Schopf, J.M. 1969. Ellsworth Mountains: position in West Antarctica due to sea floor spreading. Science, v. 164, p. 62-66.

Schwarz, E.H.L. 1896. Summary of the work done in the Tulbagh area and Worcester district. Cape of Good Hope, Geol. Comm. Ann. Rept. 1, p. 34-37.

Schwarz, E.H.L. 1897. Summary of the work done during 1897 in the Robertson, Lady Grey, Montagu, and eastern parts of the Swellendam district. Cape of Good Hope, Geol. Comm. Ann. Rept. 2, p. 51-57.

Schwarz, E.H.L. 1898a. Report on the country round Vogel Vlei, Piquetberg road and Saron. Cape of Good Hope, Geol. Comm. Ann. Rept. 3, p. 25-28.

Schwarz, E.H.L. 1898b. The country round French Hoek and Pinel. Cape of Good Hope, Geol. Comm. Ann. Rept. 3, p. 29-35.

Schwarz, E.H.L. 1905a. Geological survey of the coastal plateau in the divisions of George, Knysna, Uniondale and Humansdorp. Cape of Good Hope, Geol. Comm. Ann. Rept. 10, p. 259-290.

Schwarz, E.H.L. 1905b. Geological survey of the division of Tulbagh, Ceres, and Worcester. Cape of Good Hope, Geol. Comm. Ann. Rept. 10, p. 259-290.

Schwarz, E.H.L. 1907. The andalusite schist of George. Rec. Alb. Mus., v. 2, p. 164-175.

Schwarz, E.H.L. 1928. South African geology. Blackie & Son, London and Glasgow. 200 p.

Sclater, J.G. and Fisher, R.L. 1974. Evolution of the east central Indian Ocean, with emphasis on the tectonic setting of the Ninetyeast Ridge. Geol. Soc. America Bull., v. 85, p. 683-702.

Seward, A.C. 1914. Antarctic fossil plants: Brit. Ant. ('Terra Nova') Exped., 1910. Nat. Hist. Rept., Geol. v. 1, no. 1, Brit. Mus. (Natl. Hist.), London. 49 p.

Shand, S.J. 1917. Geology of the neighborhood of Stellenbosch. South Africa Jour. Sci., v. 15, p. 124-136.

Skeats, E.W. 1916. Report on the petrology of some limestones from the Antarctic. Brit. Ant. Exped., 1907-09, Geol. v. 2, Heinemann, London, p. 189-200.

Skinner, D.B.N. 1964. A summary of the geology of the region between the Byrd and Starshot Glaciers, South Victoria Land: in Adie, R.J., ed., Antarctic Geology, North-Holland, Amsterdam, p. 284-292.

Skinner, D.B.N. 1965. Petrographic criteria of the rock units between the Byrd and Starshot Glaciers, South Victoria Land, Antarctica, New Zealand Jour. Geology and Geophysics, v. 8, p. 292-303.

Smith, A.G., and Hallam, A. 1970. The fit of the southern continents. Nature, v. 225, p. 139-144.

Smith, D.A.M. 1965. The geology of the area around the Khan and Swakop Rivers in South West Africa. South Africa Geol. Survey Mem. 3 (South West Africa Series). 113 p.

Söhnge, P.G. 1934. The Worcester fault. Geol. Soc. South Africa Trans., v. 37, p. 253-277.

Söhnge, P.G. 1964. The geology of the Tsumeb mine: in Haughton, S.H., ed., The geology of some ore deposits of southern Africa, Geol. Soc. South Africa, p. 367-382.

Stauffer, P.H. 1967. Grain-flow deposits and their implications, Santa Ynez Mountains, California. Jour. Sed. Petrology, v. 37, p. 487-508.

Stewart, D. 1934a. The petrography of some Antarctic rocks. Am. Mineralogist, v. 19, p. 351-359.

Stewart, D. 1934b. The petrography of the Beacon Sandstone in South Victoria Land. Am. Mineralogist, v. 19, p. 351-359.

Stewart, D. 1934c. The petrography of some rocks from South Victoria Land. Am. Phil. Soc. Proc., v. 74, p. 307-310.

Stocken, C.G. 1954. The pre-Cape rocks of the central Cango. Ph.D. Dissertation, Univ. Cape Town. 75 p.

Stump, E. 1973. Earth evolution in the Transantarctic Mountains and West Antarctica: in Tarling, D.H., and Runcorn, S.K., eds., Implications of Continental Drift to the Earth Sciences, Academic Press, London-New York, p. 909-924.

Stump, E. 1974. Volcanic rocks of the Early Cambrian Taylor formation, central Transantarctic Mountains. Antarctic Jour. U.S., v. 9, p. 228-229.

Tarling, D.H. 1972. Another Gondwanaland. Nature, v. 238, p. 92-93.

Treves, S.B. 1965. Igneous and metamorphic rocks of the Ohio Range, Horlick Mountains, Antarctica. Am. Geophys. Union, Antarctic Res. Ser., v. 6, p. 117-125.

Truswell, J.F. 1970. An introduction to the historical geology of South Africa. Purnell & Sons, Cape Town. 167 p.

Truter, F.C. 1949. A review of volcanism in the geological history of South Africa. Geol. Soc. South Africa Proc., v. 52, p. 29-39.

University of Stellenbosch. 1948. Geological map of the pre-Cape rocks of the Worcester-Swellendam mountain foreland. Unpub. Univ. Stellenbosch.

van Eeden, O.R. 1970. The geology of the Republic of South Africa, an explanation of the 1:1 000 000 map, 1970 edition. South Africa Geol. Survey Spec. Pub. 18. 85 p.

Visser, D.J.L. 1957. The structural evolution of the Union. Geol. Soc. South Africa Proc., v. 60, p. 13-50.

Visser, H.N. 1967. Verspreiding en korrelasie van die formasie Klipheuwel in die Swartland en die Sandveld. South Africa Geol. Survey Annals, v. 6, p. 31-38.

Visser, H.N., and Schoch, A.E. 1973. The geology and mineral resources of the Saldanha Bay area. South Africa Geol. Survey Mem. 63. 150 p.

von Backstrom, J.W., Jansen, H., Coertze, C.B., Snyman, A.A. and Spies, J.J. 1960. Die geologie van die gebied om Nieuwoudtville, Kaapprovinsie: An explanation of Sheet 241, South Africa Geol. Survey. 49 p.

Wade, F.A. 1974. Geological surveys of Marie Byrd Land and the central Queen Maud Range. Antarctic Jour. U.S., v. 9, p. 241-242.

Wade, F.A., Yeats, V.L., Everett, J.R., Greenlee, D.W., LaPrade, K.E., and Shenk, J.C. 1965a. The geology of the central Queen Maud Range, Transantarctic Mountains, Antarctica. Antarctic Res. Rept. Ser., 65-1, Texas Tech. College. 54 p.

Wade, F.A., Yeats, V.L., Everett, J.R., Greenlee, D.W., LaPrade, K.E., and Shenk, J.C. 1965b. Geology of the central portion of the Queen Maud Range, Transantarctic Mountains. Science, v. 150, p. 1808-1809.

Walker, R.G., and Mutti, E. 1973. Turbidite facies and facies associations. Soc. Econ. Paleontologists and Mineralogists, Short Course, Anaheim, p. 119-158.

Warren, G. 1969. Geology of the Terra Nova Bay-McMurdo Sound area, Victoria Land. Am. Geog. Soc. Map Folio Ser. 12, XIII.

Webers, G.F. 1972. An unusual upper Cambrian fauna from West Antarctica: in Adie, R.J., ed., Antarctic Geology and Geophysics, Universitetsforlaget, Oslo, p. 235-238.

Wegmann, C.E. 1935. Zur deutung der migmatite. Geol. Rundschau, v. 27, p. 43-47.

Weissel, J.K., and Hayes, D.E. 1972. Magnetic anomalies in the southeast Indian Ocean. Am. Geophs. Union, Antarctic Res. Ser., v. 19, p. 165-196.

Williams, P.L. 1969. Petrology of upper Precambrian and Paleozoic sandstones in the Pensacola Mountains, Antarctica. Jour. Sed. Petrology, v. 39, p. 1455-1465.

Woodward, A.S. 1921. Fish remains from the upper Old Red Sandstone of Granite Harbour, Antarctica. Brit. Ant. ('Terra Nova') Exped., 1910, Nat. Hist. Rept., Geol., v. 1, no. 2, Brit. Mus. (Natl. Hist.) London, p. 51-62.

Yochelson, E.L., and Herrera, H.E. 1974. Un fossil enigmatico del Cambrico inferior de Argentina. Ameghiniana, v. 11, p. 283-294.

Yochelson, E.L., and Stump, E. 1977. Discovery of Early Cambrian fossils at Taylor Nunatak, Antarctica. Jour. Paleo., 51, p. 872-875.